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INTRUSIVE ROCKS OF THE OKANOGAN VALLEY AND THE PROBLEM OF THEIR CORRELATION

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ABSTRACT

Eight plutonic masses are described. These include (1) parts of three granodiorite batholiths, (2) a granodiorite stock, (3) a quartz-diorite mass probably of batholithic dimensions, (4) two small basic masses, one of metagabbro and one of malignite, and (5) a small but complex intrusion containing several alkaline types. Some of the intrusions have variable compositions, others are nearly homogeneous. Gneissic and protoclastic structures are prominent in two of the batholiths, mylonitization along shear zones in the malignite. Contact metamorphism is intense near some of the intrusions, almost absent near others. Satellitic dikes are scarce near some intrusions, abundant near others.

Cross-cutting relationships can be established for only two pairs of the intrusions. The contact between two of the batholiths is exposed but is complicated by numerous dikes, extensive deformation, and the presence of a syenitic border facies. No stratigraphic evidence is available for establishing relative ages of the intrusions. Hence the determination of genetic and temporal relationships among the intrusions must depend largely on comparisons of their chemical, petrographic, and structural characteristics.

The general problem of correlating intrusives without evidence from cross-cutting relations and age determinations is discussed in detail. Possible criteria based on (1) chemical composition, (2) petrographic characteristics, (3) structures, (4) metamorphism, and (5) dikes are tested by applying them to the Okanogan intrusives. The criteria are grouped according to their general applicability, and various combinations are used to establish relationships among the Okanogan rocks.

INTRODUCTION

The valley of the Okanogan River, in north-central Washington near the international boundary, is a region with great wealth and

variety of intrusive rocks. The intrusions range in size from batholiths to small masses barely a square mile in area, in composition from granodiorite to gabbro and malignite, in structure from crushed and foliated gneisses to rocks with normal granitic texture. Glaciation, a semiarid climate, and fairly rugged topography combine to furnish excellent exposures for detailed study.

The igneous bodies intrude the Anarchist series—a group of late Paleozoic age which has undergone considerable deformation and mild regional metamorphism. Phyllites derived from fine sediments and greenstones derived from lavas and tuffs form the bulk of the Anarchist series; quartzites, metaconglomerates, metadolerite, and fine-grained marble occur in smaller amounts. At the borders of several of the intrusives Anarchist rocks have undergone considerable contact metamorphism.

Later than the intrusives is a series of early Cenozoic sedimentary and volcanic rocks, including coarse-to-fine river deposits interbedded with andesitic flows and pyroclastics. The Tertiary rocks have been mildly deformed, and erosion has removed them from most of the area. The last chapter of the geologic history was slight erosion and abundant deposition by a Pleistocene icecap and its attendant proglacial streams.

Although debris from some of the intrusives is strangely absent from the Tertiary deposits, the intrusives were almost certainly all emplaced before these deposits were laid down. Except for bracketing the igneous rocks between late Paleozoic and early Cenozoic, fossils from the sedimentary formations give no hint as to their absolute ages. Even relative ages of the different intrusives are directly determinable in only two instances, where definite cross-cutting relations are exposed.

Because visible cross-cutting of one intrusive by another is so uncommon, their genetic relationships as well as their time relationships must remain in some doubt. The problem of correlation is especially difficult because several of the intrusives are highly variable in composition and structure. Criteria for correlation, when more definite ones are lacking, must be based on compositions, structures, and general properties of the various masses; but these criteria must be selected with extreme care, since ill-chosen ones may as effectively

distinguish between parts of a single variable intrusive as between two different intrusives. Correlation of these intrusives, therefore, becomes a problem of the selection of valid criteria from a multitude of possible ones.

One purpose of this paper is to describe the various Okanogan intrusives. A more important purpose is to discuss fully the problem

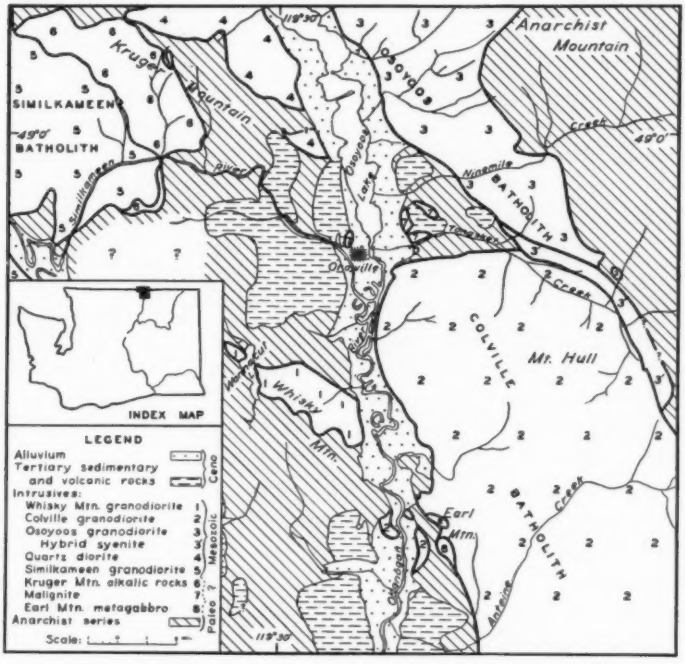

FIG. 1.—Geologic sketch map of the Okanogan Valley at the Forty-ninth Parallel

of their correlation, in the hope of drawing some useful conclusions regarding the general problem of correlating intrusive rocks when such definite criteria as cross-cutting and exact age determinations are lacking. The good exposures, the variety of intrusives represented, and the variability of individual intrusives make this an ideal area for the development and testing of different criteria.

Eight plutonic masses will be considered (Fig. 1): (1) a stock of

homophanous,[1] porphyritic granodiorite which underlies the highest part of Whisky Mountain, five miles southwest of Oroville; (2) the Colville granodiorite[2]—a highly variable mass consisting dominantly of protoclastic gneiss—the extreme northwest corner of an enormous batholith which covers much of northeastern Washington; (3) the Osoyoos granodiorite,[3] also foliated and crushed, part of a batholith which extends northward into Canada; (4) a mass of altered quartz diorite north of Oroville, which extends an unknown distance beyond the international border; (5) the Similkameen granodiorite,[4] homophanous and uncrushed, part of a batholith extending for unknown distances north and south of the border; (6) the alkaline rocks of Kruger Mountain, seven miles northwest of Oroville, which include various types of malignite and nepheline syenite; (7) a small mass of malignite which outcrops on either side of the Okanogan River near Oroville; (8) the metagabbro of Earl Mountain, a small basic mass outcropping about fifteen miles south of the international border. These eight masses are selected as convenient areal units for descriptive purposes; the actual number of separate intrusions may, of course, be greater or less than eight.

Field work occupied the summer of 1936 and parts of the two following summers. Detailed work was limited to the northwest quarter of the Osoyoos quadrangle—an area of about 200 square miles just south of the international border. Parts of the following discussion which deal with adjacent areas are based largely on the work of three other geologists: Daly, whose classic memoir on the North American Cordillera at the forty-ninth parallel (1912) includes a careful geological study of the Okanogan Valley just north of the international border; C. D. Campbell,[5] who has examined in

[1] Homophanous: "without directional structure" (J. J. Sederholm, "On Migmatites and Associated Rocks of Southwestern Finland. Part I," *Comm. Geol. Finland, Bull. 58* [1923], p. 73).

[2] J. T. Pardee, "Geology and Mineral Resources of the Colville Indian Reservation, Washington," *U.S. Geol. Surv. Bull. 677* (1918), pp. 1–181.

[3] R. A. Daly, "Geology of the North American Cordillera at the Forty-ninth Parallel," *Geol. Surv. Canada, Mem. 38* (1912), p. 455.

[4] *Ibid.*

[5] "The Kruger Alkaline Syenites of Southern British Columbia," *Amer. Jour. Sci.*, Vol. CCXXXVII (1939), pp. 527–49.

great detail the alkaline rocks of Kruger Mountain; and A. C. Waters,[6] who has studied the border of the Colville batholith south of this area.

DESCRIPTION OF INTRUSIVES

THE GRANODIORITE OF WHISKY MOUNTAIN

Outcrops of this intrusion are limited to a roughly elliptical area (long axis *ca.* N. 70° W.) of some 6 square miles in the highest part of the Whisky Mountain ridge, with a small extension west of Wannacut Lake. In spite of conspicuous microcline phenocrysts, plagioclase on the whole is considerably more abundant than potash feldspar, so that granodiorite is the most suitable name.

The rock is singularly uniform in texture and mineral content. Large (up to 4 cm.) phenocrysts of microcline appear in nearly every specimen, and the groundmass seldom changes its appearance even at contacts. Plagioclase is the most abundant light mineral, with potash feldspar a poor second and quartz in still smaller amount. Plagioclase crystals often have prominent growth lines and nearly always show faint normal zoning, more rarely oscillatory zoning; the extreme variation in composition in different specimens is $An_{21}-An_{29}$, and in a single specimen the variation seldom exceeds 4 per cent An. Narrow rims of clear albite often surround plagioclase crystals, and in some specimens tiny veinlets of albite appear. Potash feldspar (microcline and microperthite) is found mostly in the form of phenocrysts but in part as smaller crystals and as tongues and stringers of interstitial material. Boundaries between feldspar grains are mostly straight but some show a narrow strip of vermicular material which grades into the crystals on either side; the vermicular material does not show the characteristic appearance of myrmekite and probably contains little quartz. The only mafic mineral present is biotite, which occurs in aggregates of small green flakes with abundant secondary epidote (Cz_{84}), shapes of the aggregates often simulating amphibole crystals. Sphene is abundant in large, usually elongated crystals. Apatite and a little magnetite are the other accessories; no zircon was recognized (Fig. 2; Rosiwal analysis, p. 41, chemical analysis, p. 36).

[6] A. C. Waters and K. B. Krauskopf, "The Protoclastic Border of the Colville Batholith" (in press).

Locally, close to contacts, foliation and cataclastic structures are visible. But over most of the mass directional structures are entirely absent, and crushing is suggested only by wavy foliation in quartz and minor fracturing of feldspar grains. No xenoliths or schlieren were observed.

FIG. 2.—Photomicrograph, granodiorite of Whisky Mountain. K = potash feldspar; P = plagioclase; Q = quartz; S = sphene. Biotite-epidote aggregate just to left of center. Crossed nicols. ×48.

Anarchist rocks near the Whisky Mountain intrusive are riddled with dikes ranging in size from tiny stringers to large masses several hundred feet wide and in composition from aplite to quartz diorite. With the exception of some aplites these dikes are typically aschistic, their compositions varying only moderately from that of the parent intrusion. Potash feldspar appears in nearly all the dikes; plagioclase is never more calcic than sodic andesine; biotite is the only mafic mineral; epidote and sphene are abundant; albitic rims and

veinlets are often prominent. Pegmatites and lamprophyres are conspicuously absent. Quartz veins carrying small amounts of ore minerals, presumably associated with the granodiorite, are abundant on the slopes of Whisky Mountain.

The southern contact of the intrusive is mostly discordant, the northern contact concordant. Along both contacts Anarchist rocks have been mildly baked, phyllites becoming biotite hornfelses and biotite-muscovite schists. The only material added to the Anarchist rocks in large quantities is pyrite; needles of blue tourmaline were found in one locality.

No direct information regarding the age of the granodiorite is available, except that it postdates the Anarchist series. No Tertiary rocks overlie it, nor were boulders of the granodiorite found in nearby Tertiary conglomerates. Nowhere does the intrusion or its dikes make visible contact with other intrusions.

Hand specimens.[7]—Homophanous, granitic texture; light gray, speckled with black and dirty green; phenocrysts gray to pinkish gray. Plagioclase, K-feldspar, and a little quartz visible in groundmass; black spots are aggregates of small biotite flakes, often in the shape of amphibole laths, often associated with epidote; occasional sphene crystals.

Petrography.—Plagioclase crystals euhedral to subhedral, up to 1 centimeter, often partly decomposed to fine sericite, calcite, clinozoisite, with numerous inclusions. K-feldspar subhedral to anhedral, also with numerous inclusions. Biotite partly chloritized. Sphene partly altered to leucoxene. Pyrite usually present, increasing in amount toward borders.

Metamorphism.—Weak hydrothermal metamorphism suggested by partial chloritization of biotite and incipient alteration of plagioclase. Circulation of Na-rich solutions, perhaps before final solidification, indicated by albitic rims and veinlets and perhaps by vermicular margins of K-feldspar crystals. Biotite-epidote aggregates pseudomorphing amphibole possibly explained by earlier circulation of K-bearing solutions.

Structure and contacts.—Contacts sharp; no xenoliths, occasional small stringers of granodiorite invading wall rock. Often rock retains normal characteristics up to contact, sometimes lacks phenocrysts and contains abnormal amounts of quartz at contact, sometimes foliated and in one place crushed to greenish mylonite.

Dikes.—Divisible into (1) aplites, which include normal sugary aplites and

[7] Descriptive details not needed in the subsequent discussion of correlation are given in reduced type. For three of the intrusions which are fully described elsewhere these additional details are omitted.

highly siliceous rocks with fine granitic textures; (2) granodiorite porphyries, mostly rocks resembling the parent granodiorite in texture and mineral content but somewhat finer grained; (3) quartz diorites, equigranular coarse to fine-grained rocks whose abundant biotite gives them a dark color. Dikes show varying proportions of plagioclase and K-feldspar, various stages in albitization of plagioclase. Same accessories as parent mass; some aplites contain zircon, and one quartz diorite has a little blue tourmaline.

References.—Previous description of intrusion, G. O. Smith and F. C. Calkins;[8] very similar intrusion, M. S. Church.[9]

THE COLVILLE GRANODIORITE

The granitic rocks of Mount Hull and vicinity form the northwest corner of the huge Colville batholith, which covers much of the northeastern part of Washington. Since this portion of the batholith is fully described elsewhere,[10] discussion here will be limited to a résumé of features which may serve as a possible basis for correlation.

An outstanding feature of the Colville batholith is its extreme structural and petrographic heterogeneity. So great is the variation from one part of the mass to another that inclusion of all the different rock types in a single intrusion is open to question. Justification for this treatment lies in the gradational contacts between different parts of the mass and in certain petrographic peculiarities common to the diverse rock types.

For convenience of discussion this part of the batholith may be divided into (1) a zone of homophanous granodiorite, extending eastward from the highest part of Mount Hull; (2) west of this a zone 2–3 miles wide consisting dominantly of porphyritic gneiss, locally of complexly swirled gneiss; (3) farther west a wide zone of strongly banded protoclastic gneiss. Boundaries between the zones are purely arbitrary.

Zone of homophanous granodiorite.—This is a uniform, fine- to medium-grained biotite granodiorite, with plagioclase (An_{20}) slightly in excess of quartz and considerably in excess of potash feldspar.

[8] "Geological Reconnaissance across the Cascade Range near the Forty-ninth Parallel," *U.S. Geol. Surv. Bull. 235* (1904), pp. 11–97.

[9] "The Black Mountain Leucogranodiorite," *Jour. Geol.*, Vol. XLV (1937), pp. 753–74.

[10] Waters and Krauskopf, *op. cit.*

Some plagioclase crystals show faint normal zoning (extreme range 3 per cent An), a very few oscillatory zoning. Potash feldspar (mostly orthoclase) occurs in large crystals (up to 3 mm.) as well as in tiny pockets and stringers. Myrmekite is abundant, both in crush zones and along plagioclase-orthoclase boundaries, often deeply embaying the orthoclase. Biotite, the only mafic mineral, occurs as small ragged brown flakes. Muscovite is present in considerable amounts. Accessories are magnetite, apatite, a very little sphene, allanite, and zircon both as the usual stubby prisms and as long, slender rods. Crushing is not extensive but is suggested by wavy extinctions, fracturing and rounding of crystals, and slight granulations between crystals; the narrow zones of granulated material have partly recrystallized to fine aggregates of quartz, feldspar, and biotite. No epidote or amphibole was recognized.

Zone of porphyritic gneiss and swirled gneiss.—The most common rock of this zone is a biotite granodiorite–gneiss with large (up to 5 cm.) microcline phenocrysts. Proportions of the chief minerals in the groundmass are about the same as in rocks of the homophanous zone. Plagioclase is here slightly more calcic (average, An_{24}); zoning is practically absent. Quartz occurs as (1) anhedral crystals, often flattened along foliation planes; (2) very fine grains in crush zones; (3) patches of elongated, intricately intertwined crystals. An additional accessory found here as rare, tiny crystals is epidote. The gneissic foliation consists of linear structure or platy structure or both—the linear structure defined by aggregates of tiny biotite flakes or of crushed material, the platy structure largely by orientation of biotite. Strong cataclasis is suggested by large amounts of finely granulated, partly recrystallized material in irregular bands along the foliation planes.

South of Antoine Creek in this zone occur biotite-hornblende gneisses and hornblende gneisses. Commonly in these gneisses and more rarely in the biotite gneiss the foliation is swirled into ptygmatic folds.

Zone of banded gneiss.—This zone is made up of alternate light and dark bands, many of the dark bands being continuous for long distances, some of them pinching out so rapidly as to suggest badly deformed xenoliths. The difference in color between light and dark

bands is due principally to different proportions of light and dark minerals, locally to fine crushing in the dark bands. Proportions of quartz, plagioclase, and potash feldspar vary considerably, quartz and potash feldspar being more abundant in the light bands. Plagioclase is somewhat more calcic than in the other zones, averaging An_{30}. Quartz occurs in angular fragments, as tiny grains in crush zones, in veinlets, but most abundantly in pockets of clear, elongated, complexly intertwined crystals. Orthoclase occurs also as angular fragments and, in the light-colored bands, as undeformed veinlets and irregular patches. Myrmekite is always present but seldom abundant. Biotite is the commonest mafic mineral, but hornblende is abundant in some dark bands and pyroxene appears at one locality. Apatite, magnetite, allanite, and zircon are nearly always present, slender rods of zircon being especially prominent in the light bands. Sphene is more common here than in the other zones, especially in rocks containing hornblende. Epidote is rare. Chlorite and muscovite occur sporadically.

Extreme crushing in the dark bands is shown by crystal fragments set in abundant, finely granulated material. In the light bands effects of crushing have been largely obliterated by recrystallization and introduction of quartz and potash feldspar. Foliation parallel to the banding is conspicuous throughout this zone, both linear and platy structures being well developed.

Irregular patches and stringers of homophanous pegmatite are common in the two western zones, and large masses appear along the western border.

Summary of general characteristics.—Despite the seeming variety of rock types, a number of petrographic details can be set down as characteristic of the batholith as a whole. Potash feldspar (except in the pegmatite) is considerably less abundant than quartz or plagioclase and is largely orthoclase. Plagioclase varies in composition from An_{20} to An_{40}, averaging about An_{25}; zoning is absent in most crystals, faint in a few. Most slides show pockets of flamboyant quartz. Myrmekite is universal. Hornblende is abundant locally, but biotite is by far the most common mafic mineral. The accessories apatite, magnetite, allanite, and zircon are nearly always present; zircon is peculiar in its occurrence as tiny, slender rods as well

as in the usual short prisms. Sphene and epidote often appear in rocks containing hornblende but elsewhere are very scarce. Effects of crushing are apparent in nearly all slides (Rosiwal analysis, p. 41; chemical analysis, p. 36).

Linear and platy structures are well developed over much of this part of the batholith. The linear structure maintains a remarkably constant NW.-SE. trend, its extreme variation being N. 42° W.–N. 62° W. Across the top of Mount Hull the platy structure is horizontal, dipping outward at low angles around the north and west sides; in the valley of Antoine Creek is a broad anticline which plunges to the northwest near the border of the intrusion. Two joint sets are prominently developed in the batholith—one parallel or nearly parallel to the platy structure, the other a vertical set approximately at right angles to the linear structure.

Contacts of the granodiorite are sharp, in part concordant and in part discordant. Metamorphism of the intruded rocks is very slight: limestones show some recrystallization, greenstones show addition of quartz and sulphides. The only extensive metamorphism accomplished by the granodiorite was the impregnation of metagabbro on Earl Mountain with potash feldspar (p. 28).

Except for the abundant, irregularly shaped pegmatite masses, dikes definitely associated with the granodiorite in this area are limited to a few lamprophyres. Quartz veins also are singularly scarce.

Evidence from south of this area suggests that the banded gneiss of the western zone is a migmatite, derived from the incomplete assimilation and drawing-out of basic xenoliths. To explain the apparent anomaly between extensive assimilation and lack of contact metamorphism and to explain also the widespread cataclasis followed by recrystallization and by deposition of quartz and orthoclase, Waters[11] has proposed the following mechanism of intrusion: the granodiorite partially assimilated much basic material at depth, then when cooled until nearly solid moved into its present position; during the movement much of the rock was granulated, and the basic material was drawn out into bands; final solidification took place only after the batholith's emplacement in its present position.

[11] *Ibid.*

Intrusive contacts with Anarchist rocks and with the Earl Mountain metagabbro and rare boulders in Tertiary conglomerates are the only direct clues as to the granodiorite's age. The complex contact between the Colville and Osoyoos batholiths will be discussed presently.

THE OSOYOOS GRANODIORITE

For present purposes "Osoyoos batholith" will refer to the large granodiorite mass east of Lake Osoyoos, whose southern tip extends into the northern part of this area. Relations between this mass and the intrusion west of the lake will be discussed later. Along the Osoyoos-Colville contact in the Tonasket Creek Valley are some peculiar rocks of uncertain relationships, which will be discussed after the batholith itself has been described.

The total area of the Osoyoos batholith is not known. In this region outcrops occupy a roughly triangular area with its apex east of Mount Hull; Daly[12] reports a considerable widening of the batholith to the north. The composition of the rock lies approximately between that of granodiorite and quartz diorite, but granodiorite is the more suitable term for most of the specimens collected. Foliation and cataclasis are almost as conspicuous and as widespread as in the Colville batholith.

The granodiorite is fairly uniform in texture and composition except near its borders. Amounts of quartz and plagioclase are nearly equal, and potash feldspar is always distinctly subordinate. Quartz occurs (1) as anhedral crystals, often elongated with the foliation, (2) as tiny grains in partly recrystallized crush zones, (3) as inclusions in plagioclase, and (4) in conspicuous pockets of clear, intricately intertwined crystals like those so common in the Colville batholith. Plagioclase (An_{38}; composition nearly constant, except for more sodic varieties near contacts) occurs as anhedral crystals, usually with rounded or broken edges; zoning is nearly absent. Potash feldspar is mostly orthoclase, commonly in irregular patches and tiny veinlets. Myrmekite is usually present in small amounts. Biotite, hornblende, and epidote commonly occur in platy aggregates of indefinite shape. Hornblende is conspicuously variable, sometimes nearly equal to biotite in amount, sometimes completely

[12] *Op. cit.*, map-sheet 11.

absent. Properties of the hornblende are somewhat unusual: n = *ca.* 1.685; birefringence *ca.* 0.017; $c \wedge Z = 20°$; $2V$ = *ca.* 55°; strongly pleochroic from light yellow-brown to green to blue-green. Epidote (Cz_{78}) is nearly always present in considerable quantity, both in veinlets and scattered through the rock. The usual accessories include magnetite, apatite (scarce, in tiny crystals), zircon (in stout prisms), and sphene (more abundant near contacts). Allanite is reported by Daly from north of the international boundary but was not seen in any of the slides examined. Secondary minerals (calcite, chlorite, sericite, epidote in feldspars) are scarce except near contacts. Crush material is nearly always present, mostly as thin streaks along crystal margins; no specimen shows the extreme crushing found along the west margin of the Colville batholith. Foliation is defined by alternation of bands rich in quartz and in feldspar and by the platy aggregates of dark minerals (Fig. 3; Rosiwal analysis, p. 41).

Near the contacts composition and texture are highly variable. Plagioclase commonly becomes more sodic, and potash feldspar appears in larger amounts. Some specimens lack biotite, others lack hornblende. Some specimens show allotriomorphic textures as if they had completely recrystallized. Foliation is sometimes so pronounced that the rock resembles a mica schist. Hydrothermal alteration locally gives greenish rocks whose thin sections show saussuritized feldspars and chloritized mafics.

Platy foliation is well developed over most of the batholith. Its strike varies from N. 20° W. to N. 70° W.; the dip varies from 20° E. to 70° E. and according to Daly becomes steeper to the north. The average attitude is similar to that of the platy structure in the Colville batholith on the north side of Mount Hull. Easily visible linear structures are limited to isolated localities near the contacts. A few well-defined xenoliths and larger, more indefinite masses of basic gneiss appear sporadically along the contact but show no conspicuous orientation. The most prominent joints lie in a vertical plane trending roughly N. 60° E.

The Anarchist contact is for the most part concordant, locally sharply discordant. At the few places where it is well exposed the contact seems to be gradational, high-grade metamorphic rocks

passing into strongly foliated gneisses by imperceptible steps. The metamorphic aureole in the Anarchist rocks, out to the lower side of the biotite zone, varies in width from $\frac{1}{4}$ to $\frac{1}{2}$ mile. Near the batholith metamorphism is extreme, phyllites becoming amphibole gneisses and diopside gneisses, greenstones becoming amphibole

FIG. 3.—Photomicrograph, Osoyoos granodiorite. Note feldspar fragments with jagged edges and wavy extinction, stringers of finely crushed material, and large patch of flamboyant quartz. Crossed nicols. ×18.

rocks resembling diorite. Additive metamorphism has played a minor role, some of the higher-grade types containing small amounts of added quartz and orthoclase and occasionally tourmaline.

Dikes, both in the granodiorite and in the bordering rocks, are very numerous along the Osoyoos contact. These are typical diaschistic dikes, aschistic types like those on Whisky Mountain being conspicuously absent. The dikes are conveniently divided into three

groups: (1) pegmatites, of which the most abundant variety is a coarse-grained rock containing albite, microcline, quartz, muscovite, and pink garnet, the white feldspars being set off conspicuously by blue-gray quartz; (2) aplites, for the most part of normal, sugary texture; (3) lamprophyres, which include basic dikes with a great variety of compositions and textures. Quartz veins of all sizes are locally abundant in the metamorphics adjacent to the batholith.

The age of the batholith is bracketed between the Anarchist rocks and the Tertiary by the intrusive contact with the former and by the presence of unmetamorphosed andesite and sandstone within a few hundred feet of the granodiorite margin. Intrusion postdating the Oroville malignite is suggested by xenoliths of malignitic composition (p. 26). Age relations with the Colville batholith will be discussed below.

Hand specimens.—Light gray, with platy aggregates of dark minerals; often a greenish cast because of abundant epidote; joint surfaces conspicuously green from epidote. Medium grained, usually well foliated. Quartz, feldspar, biotite, epidote easily visible, usually hornblende and sphene also.

Petrography.—Texture variable, most commonly anhedra of plagioclase and quartz (*ca.* 1 mm.) in matrix of finely granulated material with recrystallized quartz and K-feldspar. Plagioclase usually clear, larger crystals sometimes contain a few tiny grains of epidote. Myrmekite in crush zones and at edges of orthoclase crystals. Biotite in green to brownish-green flakes of all sizes, the smaller ones scattered along foliation planes, the larger ones in aggregates with hornblende and epidote. Part of the epidote in large crystals scattered through rock; has appearance of primary mineral, at least in part earlier than biotite.

Metamorphism.—Undeformed quartz and K-feldspar suggest that here, as in the Colville batholith, deformation preceded final crystallization; this is supported by the swirling of the foliation around dikes, described below. The relation of cataclasis to foliation is not altogether clear, since streaks of granulated material do not so obviously parallel the foliation as in the Colville batholith; but absence of extensive recrystallization of the granulated material, on the one hand, and lack of distortion or granulation of quartz and biotite (which largely determine the foliation), on the other hand, suggest strongly that gneissic and cataclastic structures were developed by the same deforming stress.

Daly has postulated profound metamorphic changes in the rocks north of the international border. During granulation of the rock he supposes its minerals to have been partly dissolved by hot solutions, the minerals most readily affected being hornblende, biotite, magnetite, apatite, and sphene. The dissolved material was in part redeposited as fresh crystals of biotite and epidote, in part

concentrated in zones of maximum shear to form an unsheared basic hornblende gneiss. With this ingenious hypothesis of leaching and redeposition Daly accounts for the observed variations in texture and in the proportions of biotite, hornblende, epidote, and sphene.

Three bits of evidence from this area seem not in accord with Daly's hypothesis: (1) the accessories magnetite, apatite, and sphene are no more abundant in rocks containing hornblende than in those with biotite as the only mafic mineral; (2) according to Daly's idea the rock should be markedly heterogeneous, with biotite and epidote concentrated along lines of easiest escape, but here the rock is conspicuously uniform over broad areas; (3) so extensive a redistribution of material could hardly have left plagioclase crystals with jagged edges, with no alteration, and almost no zoning. These arguments do not disprove Daly's hypothesis, but it seems simpler to ascribe the different rock types to original variations in composition and to suppose that fresh biotite and epidote were deposited in the last stages of crystallization without any large-scale migration of material.

Whether the diversity of rocks along the batholith's margin is best explained by assimilation of foreign material, by earlier formation of a basic border, or by local concentration of solutions from the cooling magma cannot be answered satisfactorily. Gradational contacts with Anarchist rocks certainly suggest the possibility of assimilation, but xenoliths are small in size and number.

Structure.—Within a mile or so of the eastern contact the foliation is often somewhat swirled, locally is absent altogether. Immediately adjacent to a concordant contact the foliation usually parallels the contact but sometimes shows erratic changes of direction; at a discordant contact the foliation usually parallels the structure of the adjacent metamorphics. The general absence of linear structures, the frequent failure of platy structures to follow contacts, the gradational nature of the exposed contacts, all suggest that foliation in the Osoyoos batholith is not directly related to large-scale translational movement but is due to partial granulation and recrystallization in place. This conclusion, if valid, is in marked contrast to that suggested for the Colville batholith.

Small-scale structures associated with small aplite and lamprophyre dikes give striking evidence for the hypothesis that deformation occurred while the granodiorite was in a semiplastic state. A few of these dikes have been broken, and their fractured ends are displaced a few centimeters. In the fracture is ordinary gneissic material, its foliation somewhat swirled but continuous with that in the gneiss on either side. Evidently at the time the dike was intruded and solidified, the gneiss was still sufficiently fluid to heal a fracture. Thus, deformation of at least this part of the gneiss must have occurred when it had a consistency like that of pitch, solid enough to maintain cracks for the intrusion of dikes but still capable of slow fluid motion.

Dikes.—Pegmatite occurs in small dikes (seldom over 30 cm. wide) and in large irregular masses along the batholith's margin. The small dikes have a char-

acteristic structure: fine-grained material near the edges, large feldspar crystals in blue-gray quartz in the center, a row of tiny pink garnets a centimeter or so from either wall. A rarer type of pegmatite consists almost entirely of pink microcline and microperthite crystals. Aplite dikes are typically small, following erratic courses through the granodiorite and adjacent rocks. The most numerous dikes are the lamprophyres. These range in size from tiny stringers to massive dikes 10 meters wide; the smaller ones show no favored orientation, but the larger ones usually parallel the principal joint set. A great variety of textures and compositions is represented, but the majority consist largely of euhedral to subhedral andesine and hornblende; biotite and epidote are often present, K-feldspar and quartz more rarely. A few dikes of every class show marked shearing.

References.—Description of granodiorite north of the international border, Daly;[13] discussion of biotite-epidote associations, L. Duparc[14] and F. C. Phillips;[15] description of foliation around fractured dikes similar to those found here, J. J. Sederholm[16] and R. Balk and F. F. Grout.[17]

The Osoyoos-Colville border zone.—Although the zone where these two large granitic masses meet is well exposed, a study of the contact helps little in clarifying their relationships. The chief difficulty is the presence along the contact of peculiar rock types which cannot with certainty be assigned to either batholith. Since the rocks of this contact zone are fully described in another paper,[18] the present discussion is limited to those characteristics which have a bearing on the problem of correlation.

Rocks of the border zone are gneissic hornblende syenites, cut by many lamprophyre dikes and by innumerable small and large masses of pegmatite. The syenites extend around the border of the Colville batholith in a half-mile-wide strip from the valley of Tonasket Creek to a point due east from the summit of Mount Hull. Basic syenites

[13] *Ibid.*

[14] "Sur l'origine de l'épidote dans certains granits," *Compt. rend. Soc. phys. Genève*, Vol. XXXVIII (1921), pp. 100–102.

[15] "Petrographic Notes on Three Rock-Types from the Shetland Islands," *Geol. Mag.*, Vol. LXV (1928), pp. 500–507.

[16] "On Migmatites and Associated Rocks of Southwestern Finland. Part II," *Comm. Geol. Finland, Bull. 77* (1927), pp. 1–140.

[17] "Structural Study of the Snowbank Stock," *Bull. Geol. Soc. Amer.*, Vol. XLV (1934), pp. 621–36.

[18] Waters and Krauskopf, *op. cit.*

predominate at the western end of this strip, siliceous varieties toward the eastern end.

Potash feldspar is more abundant in the syenites than in either batholith, quartz much less abundant. Plagioclase ($An_{16}-An_{18}$) is somewhat more sodic than in the adjacent granodiorites. Hornblende is the only mafic mineral. Epidote is uniformly present but abundant only in occasional large clots and veinlets. Accessories are abundant, including garnet in the more basic varieties. Textures range from normal granitic texture to the mosaic texture of metamorphic rocks and a strongly cataclastic texture approaching that of mylonite.

The most prominent feature of the border zone is its extremely complex structure. Foliation in the basic varieties is fairly even, but the rock is cut into great angular chunks by pegmatite of the quartz-poor variety. Foliation in the siliceous syenites is intricately swirled; pegmatite stringers and intertwining minor zones of shear wind erratically through the rock. Large shear zones of several ages are represented—some completely recrystallized, some lined with scarcely recrystallized mylonite, some with prominent slickensides.

At the one locality where a syenite-Osoyoos contact is exposed—in the valley of Tonasket Creek—it appears to be gradational. Elsewhere the border zone is separated from Osoyoos granodiorite by a narrow strip of highly metamorphosed Anarchist rocks. Intervening pegmatite prevents actual contact between syenite and Colville granodiorite, but the two approach each other closely in several places. There is no evidence that one rock was markedly more fluid than the other at the time of intrusion—no chilled borders, no flow structures oriented with the contact, no angular xenoliths, no cross-cutting dikes chilled more in one mass than in the other. Thus contact relations suggest that the syenite is not markedly different in age from the batholiths on either side and hence that the batholiths themselves are approximately contemporaneous.

Whether the syenite is genetically related to one batholith rather than to the other must remain an unsettled question. Field relations of the syenite, the sodic character of its plagioclase, and its abundant potash feldspar suggest kinship with the Colville granodiorite. On the other hand, at least five lines of evidence point to closer ties with

Osoyoos rocks: (1) hornblende in the syenite has the same peculiarities as that of Osoyoos granodiorite (p. 13); (2) epidote and sphene are common in the syenite; (3) zircon in the syenite is never found as slender rods; (4) other rocks along the Osoyoos border (e.g., near Osoyoos Lake) resemble the basic syenites; (5) the syenite shows strong foliation similar to the foliation in adjacent parts of the Osoyoos batholith, while the Colville batholith is here nearly structureless.

Perhaps (although this hypothesis is far from proved) the peculiarities of the syenite can be best explained by assuming that the rock is a hybrid, formed by the action of emanations from both batholiths on Anarchist rocks caught between them. Evidence has been presented that the two batholiths were subjected to deforming stress in the last stages of crystallization and that the Colville batholith at least moved a considerable distance in a partly crystallized condition. Evidence from contacts with the syenite suggests strongly that solidification of the batholiths took place at about the same time. It seems reasonable to suppose that two semiplastic masses undergoing deformation, with one moving against the other, would produce just such a contact zone as the strip of hornblende syenite—an intensely deformed zone in which heat and volatiles would be concentrated sufficiently to make over fragments of country rock and to riddle the surroundings with dike material at various stages of the cooling.

THE QUARTZ DIORITE NORTH OF OROVILLE

The granitic rocks west of Osoyoos Lake, which Daly[19] mapped with the Osoyoos batholith, extend northward for an unknown distance along the Okanogan Valley. Only a small area of these rocks near the international border will be discussed here. The dominant rock type is probably best called "quartz diorite," although extensive hydrothermal alteration makes the identification not altogether certain.

The prevailing green cast of these rocks betrays the alteration of their mafic minerals to chlorite and epidote, of their feldspars to albite (An_{10}) studded with fine sericite, calcite, and epidote. No

[19] *Op. cit.*, p. 439.

original plagioclase was found in any of the slides examined. Some of the old feldspar-shapes are rimmed with clear albite. Potash feldspar is scarce, myrmekite is absent. Quartz occurs as large anhedra. No biotite was found except as chlorite pseudomorphs. Pseudomorphs of amphibole are common, but no original hornblende was found except in a fine-grained dioritic facies; here its properties are those of normal hornblende, not the unusual variety found in the Osoyoos granodiorite. Epidote (Cz_{83}) is abundant, in large crystals, in aggregates of small crystals, and along veinlets. Sphene, magnetite, and apatite are the common accessories; no zircon or allanite was observed. Pyrite and other sulphides often occur in veinlets, seldom in the rock itself.

No foliation or intimate cataclasis was observed, but the rock is rudely schistose along numerous zones of strong shear. Daly remarks for parts of the mass farther north that foliation west of Osoyoos Lake has the same attitude as that to the east but is much less conspicuous. Considerable contact metamorphism by the quartz diorite is suggested by the development of secondary hornblende at greenstone contacts and of garnet rock from limestone. A few aplite, pegmatite, and lamprophyre dikes are associated with the quartz diorite, the lamprophyres showing the same extensive chloritization and epidotization as the parent rock. Quartz veins, some carrying conspicuous amounts of sulphide minerals, are common in the surrounding rocks and in shear zones within the quartz diorite.

Megascopically and microscopically the quartz diorite resembles the hydrothermally altered rocks occasionally found along the margin of the Osoyoos batholith, but it differs markedly from normal Osoyoos types. Direct evidence regarding its age is limited to its intrusive contact with Anarchist rocks.

Hand specimens.—Dirty light-green and gray-mottled quartz often showing a violet tinge; finer-grained types fairly dark green. Uniform, medium-fine to medium-coarse grained; homophanous. Feldspars cloudy, yellowish; much green chlorite, often in recognizable amphibole shapes.

Petrography.—Quartz abundant in coarse-grained types, scarce or absent in fine-grained types. K-feldspar in fine-grained types as clear, irregular patches and veinlets. Sphene sometimes abundant, often in large crystals.

Structure and metamorphism.—The fine-grained, dioritic facies is restricted to an area of half a square mile straddling the international border. The diorite

shows considerable textural and mineralogical variety, but exposures are insufficient to reveal much about the relations of the different varieties to one another or to the quartz diorite. At least three possibilities regarding the origin of the diorite are open: (1) it may be an Anarchist metadolerite unrelated to the quartz diorite; (2) it may be greenstone highly metamorphosed by the quartz diorite; or (3) it may be a border facies of the quartz diorite. The last alternative seems most reasonable because of the sharp boundaries of the diorite against adjacent greenstones and because of its petrographic relationships with the quartz diorite.

THE SIMILKAMEEN GRANODIORITE

According to Daly, the Similkameen batholith is about 14 miles wide where it crosses the international border. Southward it extends for at least 20 miles from the border, probably much farther; its northward extent is wholly unknown. Since time did not permit an adequate study of the mass, the following remarks are compounded from the work of Daly[20] and Campbell,[21] from information supplied by Waters, and from a reconnaissance along a part of the east side of the batholith.

The batholith consists for the most part of uniform, homophanous granodiorite, allied to the alkaline granites (Daly). Some parts have large potash feldspar phenocrysts, and somewhat more basic varieties occur in places on the eastern margin. The staple rock has only slightly less potash feldspar than plagioclase, with quartz somewhat subordinate. Plagioclase is characterized by strong progressive zoning, from An_{40} to An_{25}; some crystals have clear rims of more sodic material. A few show pronounced oscillatory zoning. (Daly does not mention zoning and gives as average composition An_{30}.) Potash feldspar is largely microcline and microperthite in large crystals and in irregular patches between other minerals. Plagioclase shows irregular somewhat lobate margins toward potash feldspar, but myrmekite was observed only as one small patch on one slide; neither Daly nor Campbell mentions intergrowths of any sort. Biotite and hornblende occur in roughly equal amounts: biotite in large, greenish-brown flakes and in small crystals along hornblende cracks; hornblende in fairly large, rather light-green crystals. Epidote appears in small amount, often along cracks in biotite or hornblende.

[20] *Ibid.*, pp. 455–59.

[21] *Op. cit.*, p. 538.

Common accessories are sphene, apatite, magnetite. Biotite is slightly altered to chlorite, plagioclase to clinozoisite and sericite. Specimens collected near the border of the mass commonly differ from this description in their lower amounts of quartz and biotite, their higher potash feldspar and amphibole, the presence of augite

FIG. 4.—Photomicrograph, Similkameen granodiorite. K = microperthite; P = plagioclase; Q = quartz; B = biotite. Note zoning in plagioclase and absence of cataclastic structure. Crossed nicols. ×26.

in small amounts, and the addition of zircon as an accessory. No specimens show cataclasis or foliation (Fig. 4; Rosiwal analysis, p. 41).

At its contacts, which are usually sharp and often discordant, the batholith bakes Anarchist phyllites, quartzites, and conglomerates to hard, black hornfelses.

Regarding the number and kinds of dikes associated with the Similkameen granodiorite, reports from different observers are con-

flicting—probably because each was concerned with a different part of the batholith. Near the international border Daly mentions "a few basic dikes," two of which he describes more explicitly as "medium-grained hornblende-diorite porphyrite" and "fine-grained augite-hornblende porphyrite." Near the contact with the Kruger Mountain alkaline rocks, north of the border, Campbell describes (1) four porphyritic dikes with phenocrysts of oligoclase in a groundmass of microcline and badly altered hornblende; (2) six lamprophyres; (3) pegmatites and aplites, of unspecified size and number, with compositions "the same as that of most such granitic satellites."[22] Waters, from observations on Aeneas Mountain, about 20 miles south of the border, reports numerous dikes along the contact —microdiorites, granodiorite porphyries, and diorite porphyries. From these descriptions Similkameen dikes include both aschistic and diaschistic types. Quartz veins, some with traces of sulphide minerals, are common near the batholith.

The granodiorite cuts Anarchist rocks, and boulders from it are abundant in the Tertiary conglomerate. Its contact relations with the Kruger Mountain syenites will be discussed below.

THE ALKALINE ROCKS OF KRUGER MOUNTAIN

Outcropping in an area of about 12 square miles in the northwest corner of the mapped region (Fig. 1) is a complex mass of alkaline rocks described by Daly and in more detail by Campbell. The following summary is taken from the work of these authors.

Campbell divides the principal rocks present into four groups: malignites (most abundant), nepheline syenites, femic nepheline syenites, prophyritic alkaline syenites. Only the first two require attention here.

The malignites are medium-dark to dark rocks of granitic texture, resembling basic syenites or diorites. Essential minerals of the malignites are microcline, augite, and hastingsite; others usually present include orthoclase, oligoclase, biotite, nepheline, and melanite. Microcline occurs in crystals of all sizes up to large phenocrysts, usually rimmed and often invaded by albite. Nepheline occurs in tiny round grains enclosed in feldspar and as interstitial material be-

[22] *Ibid.*

tween feldspar crystals; much of the interstitial nepheline has been altered to a fibrous mixture of sericite and feldspar. Plagioclase is always a minor constituent, its composition showing considerable variation (An_{13}—An_{25}). The augite is nonpleochroic and shows an unusually high extinction angle (60°); it is almost invariably rimmed with hastingsite, which also occurs in separate crystals, frequently with feldspar inclusions. Biotite appears in small amount, usually associated with pyroxene and amphibole. Garnet forms small light-colored crystals and larger brown ones often enclosing feldspars and accessories. Accessories include sphene, apatite, and magnetite—the latter often rimmed with sphene granules. Secondary epidote, chlorite, and calcite appear in some slides. The rock shows some crushing, and foliation is developed locally near contacts.

The nepheline syenites occur in small, irregular masses and as apophyses invading malignite; contacts suggest that the syenite invaded the malignite at a time when the latter was still semifluid. Microcline is the chief mineral of the syenites, often the only feldspar. Nepheline occurs in the same manner and with the same alteration products as in the malignites. Augite, hastingsite, and biotite appear in minor amounts as tiny grains. Garnet and apatite are fairly abundant, sphene and magnetite scarce. The nepheline syenites are in part foliated, in part homophanous.

Contacts of the alkaline rocks with the Anarchist series are sharp, except locally where small greenstone xenoliths have been caught up, softened, and stretched by the magma. No metamorphic changes are observable in the Anarchist rocks except, at one locality, the addition of pink feldspar metacrysts to greenstone.

The Similkameen batholith is the only intrusive mass with which the alkaline rocks make visible contact. For most of its length this contact is gradational, with a transition zone about 200 yards wide; but in two places the intrusion of granodiorite into malignite is established by plutonic breccias with angular malignite blocks separated by aplite. Campbell argues at some length the question of whether the alkaline rocks should be considered a border facies of the batholith or a separate intrusion invaded by the batholith but does not reach a definite conclusion.

THE MALIGNITE NEAR OROVILLE

On either side of the Okanogan Valley just north of Oroville are outcrops of a dark, coarse-grained rock which in thin section closely resembles the malignites of Kruger Mountain. Potash feldspar, augite, and hastingsite are its chief minerals, and patches of fine-grained muscovite and feldspar may represent decomposition products of nepheline. Despite the absence of certainly identified nepheline, the rock is so similar to some varieties described by Campbell that "malignite" seems the most appropriate name.

In composition and texture the rock is fairly uniform, specimens from east of Oroville generally showing a somewhat smaller grain size and a slightly higher percentage of light minerals. The dominant light mineral is potash feldspar (mostly microcline and microperthite) in large anhedral to subhedral crystals. Sodic andesine (An_{32}) is present in small, partly decomposed grains, often surrounded by rims of clear albite. Albite also occurs in irregular blobs and as rims around potash feldspar. Nearly colorless augite rimmed with dark-green hastingsite is always present. Apatite, magnetite, sphene, and epidote are the common accessories, grains of magnetite being characteristically rimmed with sphene. Greenish-brown biotite, muscovite, calcite, and chlorite appear in small and variable amounts. Distributed irregularly throughout the malignite are small dikes and patches of coarse-grained, pinkish-gray syenite—a rock consisting almost entirely of feldspar with a very little augite and scattered black garnets.

The malignite shows no foliation and little of the intimate crushing between mineral grains so characteristic of the Osoyoos and Colville granodiorites. Cataclastic structures of another sort, however, are well developed. Along shear zones the malignite is locally powdered into a purple-brown mylonite, which sometimes appears as thin, sharp-walled streaks through the rock, sometimes as the matrix for coarse cataclastic breccias. Outside the brecciated zones, even close to their edges, the malignite shows only minor crushing (Figs. 5 and 6).

Contacts of the malignite on the hill east of Oroville show clearly that it is intrusive into Anarchist rocks and is overlain by Tertiary conglomerates. Concerning its relations with other intrusions there

are two conflicting lines of evidence. (1) The lack of foliation and intimate crushing suggests that the malignite is later than either of the adjacent large granodiorite bodies. (2) On the international border just east of Osoyoos Lake are a few small basic xenoliths in the

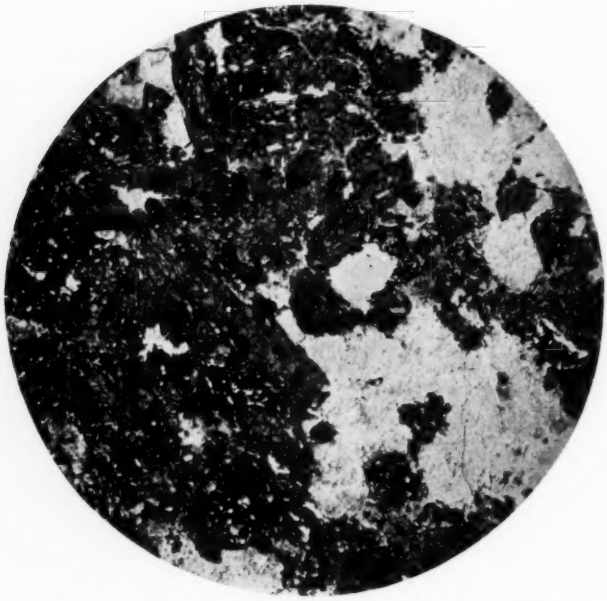

FIG. 5.—Photomicrograph, malignite from near Oroville. Hastingsite (dark gray) rims surround and penetrate augite cores; black magnetite grains show narrow rims of sphene. Light patches mostly microperthite. Parallel light. ×18.

Osoyoos batholith which in thin section show a startling resemblance to the malignite. The xenoliths have a higher proportion of mafic minerals than the malignite and apparently do not contain albite; but thin sections show the same colorless pyroxene surrounded by hastingsite, the same large crystals of microcline, the same abundant, good-sized prisms of apatite, the same magnetite grains rimmed with sphene. It seems most reasonable to consider these xenoliths as fragments of the malignite and to conclude, therefore, that the malignite is older than the adjacent batholiths. The foliation and intimate

crushing in the batholiths are probably of protoclastic origin, and absence of these structures in the malignite may reflect simply its smaller size and lower temperature during intrusion of the batholiths rather than a younger age.

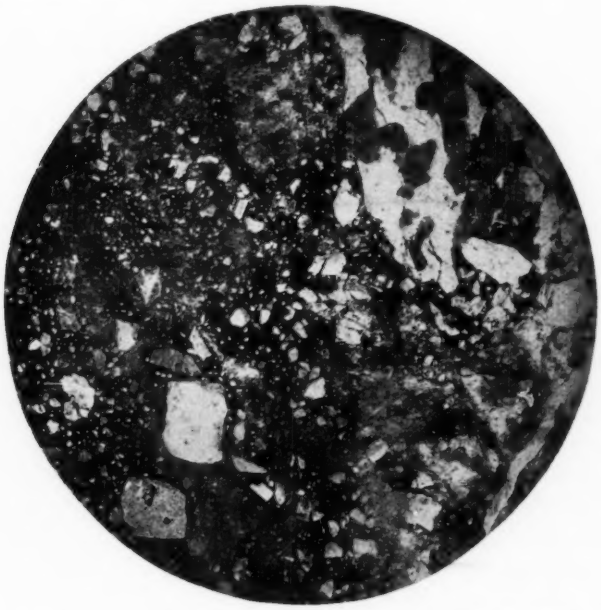

FIG. 6.—Photomicrograph, mylonitized malignite. Part of shear zone with streaked mylonite at left, fractured malignite at right. Parallel light. ×18.

Hand specimens.—Medium-coarse grained, green-black and light pinkish-gray spotted, dark minerals somewhat in excess of light. Feldspar and pyroxene recognizable.

Petrography.—Granitic texture. K-feldspar and albite fresh, plagioclase somewhat decomposed to calcite and sericite. Augite has unusually high extinction angle ($c \wedge Z = ca.\ 59°$). Hastingsite, both as rims around augite and as separate crystals, often skeletal with inclusions of feldspar.

Metamorphism.—Occurrence of hastingsite as rims around pyroxene suggests its formation in a late magmatic stage. Abundance of albite in irregular patches and as rims around other minerals probably means addition of Na after solidification. Perhaps (as suggested by Campbell for the Kruger Mountain rocks)

Na-rich solutions accomplished both introduction of albite and conversion of hornblende to hastingsite.

Exomorphism.—Contacts poorly exposed. No indication of metamorphism except hydrothermal leaching along probable fault contact.

Form and contacts.—Small oval hill west of Oroville has no exposed contacts. Malignite east of Oroville has north contact dipping moderately south, south contact in part flat, in part dipping north; hence has form of large wedge with east-west axis. Contacts with Anarchist in part concordant, in part discordant.

Structure.—Numerous shear zones, especially in malignite east of Oroville. Rock along shear zones in all stages of brecciation and mylonitization. Mylonites very fine, opaque in thin section, with scattered angular feldspar fragments. No sign of recrystallization in mylonite except stray wisps of chlorite.

THE METAGABBRO OF EARL MOUNTAIN

Basic intrusive rocks outcrop in two elongated masses on the two peaks of Earl Mountain and in numerous dikes cutting Anarchist phyllites and quartzites both on Earl Mountain and on the hills directly across the Okanogan Valley. The petrographic features of these rocks seem best explained as the result of the complex metamorphism of a gabbro, so that "metagabbro" seems the most appropriate name for the mass as a whole.

The rock of freshest appearance, both in hand specimen and in thin section, is an undoubted gabbro, consisting of labradorite and hornblende with the usual accessories. The light-green color, slight pleochroism, and frayed, lathlike forms of the hornblende suggest its secondary origin, but other signs of metamorphism are absent. This type is rare; most specimens show considerable metamorphism, in part the usual saussuritization and chloritization, in part an invasion of the rock by potash feldspar derived presumably from the adjacent Colville granodiorite. The chief evidence that potash feldspar has been added to the rock, instead of being an original constituent (as in the malignites of the Kruger Mountain and Oroville districts), is as follows: (1) the occurrence of potash feldspar in irregular masses rather than well-formed crystals, especially its occurrence in veinlets, pockets, and stringers cutting other minerals (Fig. 7); (2) the variation in potash-feldspar content with distance from the visible Colville contacts—far from the contact is nearly unaltered gabbro, while along the contact is a hybrid rock with characteristics of both pegmatite and metagabbro; (3) the absence of unusual min-

erals characteristic of the alkalic rocks to the north, such as hastingsite, nepheline, melanite.

Some of the metagabbro dikes are markedly discontinuous, and isolated fragments are often found surrounded by Anarchist rocks,

FIG. 7.—Photomicrograph, metagabbro from Earl Mountain. *K* = potash feldspar; *P* = plagioclase (studded with epidote); *C* = chlorite. Note irregular patches of potash feldspar, especially narrow tongue penetrating plagioclase just above center. Crossed nicols. ×48.

suggesting that the dikes have been badly fractured; but the metagabbro shows little of the cataclastic structure so conspicuous in the near-by batholith. Gneissic foliation is well developed in some of the dikes, but rocks from the larger masses are strictly homophanous.

That the metagabbro postdates the Anarchist series is proved by clear-cut intrusive contacts, that it antedates the Colville mass by

pegmatite dikes cutting metagabbro and by widespread addition of potash feldspar. Possibly the intrusion is genetically related to the spilites, albite dolerites, and chlorite schists which are abundant in the upper part of the Anarchist series.

Hand specimens.—Mottled, dirty green, homophanous (except in some dikes), medium to coarse, uniform grained to slightly porphyritic, dark minerals forming half or more of the visible surfaces. White, cloudy feldspar and dark-green, more or less chloritized hornblende nearly always visible; pale yellow-green epidote frequently visible; quartz and purple-brown biotite rare.

Petrography, Spec. 1.—Interlocking feldspar crystals (up to 3 mm.), aggregates of chlorite and hornblende. Chief feldspar is K-feldspar (microcline, microperthite, and probably orthoclase), in irregular cracks and patches, often surrounding other minerals. Plagioclase is albite (An_{10}), small crystals often clear, large ones cloudy and filled with epidote (Cz_{86}). A little dark-green, strongly pleochroic hornblende, always surrounded by chlorite. Accessories are sphene, ilmenite, apatite, a little pyrite. Considerable calcite. No quartz or biotite.

Spec. 2.—Needles and laths of hornblende on a fine groundmass of feldspar grains (mostly 0.05–0.2 mm.). Chief feldspar albite (An_{08}), in rough mosaic texture, studded with tiny calcite and epidote grains. K-feldspar in small, clear grains and irregular patches. A little quartz, considerable biotite. Biotite and hornblende somewhat altered to chlorite. Considerable epidote (Cz_{86}). Accessories are sphene, pyrite, magnetite, apatite. Considerable calcite.

Spec. 3.—Bent and frayed hornblende needles and plates with light-brown biotite in small flakes on swirled groundmass of feldspar. Chief feldspar andesine (An_{36}), in crystals of all sizes, filled with tiny clinozoisite (Cz_{94}) grains; much andesine as very fine material swirled between other crystals, perhaps crush material recrystallized to a fine mosaic. Some biotite apparently growing in cracks in hornblende. K-feldspar rare, in tiny patches and as a small veinlet. Accessories are sphene, ilmenite, apatite. No quartz or chlorite identified. Wavy foliation.

Metamorphism.—Albite, epidote, clinozoisite probably produced by saussuritization of more calcic plagioclase; no good evidence for addition of albite. K-feldspar introduced from Colville batholith. Feldspar mosaics as in Specimen 2 above show extensive recrystallization but appear only locally. Possibly biotite in part formed by reaction between K-rich solutions and hornblende. Partial chloritization, addition of pyrite and probably of quartz are minor metamorphic changes. Hybrid rocks found at the metagabbro-pegmatite contact on the north peak contain much quartz in fine grains, no hornblende, abundant, partly chloritized, dirty-green biotite, Fe-rich epidote (Cz_{80}), a little muscovite, abundant pyrite, sphene, and apatite. This type grades into both pegmatite and metagabbro.

Exomorphism.—Phyllites and quartzites commonly metamorphosed to biotite schists and hornfelses, greenstones to fine amphibole schists. Local higher-grade metamorphism: phyllites become garnet-mica schists and hornblende-mica schists; greenstones become fairly coarse amphibole rocks.

Contacts and structures.—Contacts of larger masses discordant; dikes follow strikes of adjacent phyllites approximately. Innumerable joints and small faults; most prominent joints N. 50° E. vertical. Cataclastic structure local, not conspicuous. Foliation only in some dikes.

References.—Previous description of intrusion, A. F. Woodward.[23] Similar metamorphism of basic rocks by siliceous magmas reported by J. J. Sederholm,[24] A. K. Wells and S. W. Wooldridge,[25] S. R. Nockolds,[26] H. H. Thomas and W. C Smith,[27] J. H. Taylor and E. A. Gamba,[28] J. Jung,[29] and W. A. Deer.[30]

CORRELATION OF INTRUSIVES

The diversity of the Okanogan intrusives in composition, in structure, in metamorphic activity, and in associated dikes has been sufficiently emphasized by the descriptions of preceding paragraphs. The few definite bits of evidence regarding age relationships have been mentioned—the intrusion of Earl Mountain metagabbro by Colville pegmatite, the intrusion of malignite by Similkameen granodiorite on Kruger Mountain, the complex Osoyoos-Colville contact suggesting approximate contemporaneity for these intrusions, the probable xenolith of malignite in Osoyoos granodiorite. Three suggestions of Daly's regarding correlation have been noted: a possible relationship between the quartz diorite north of Oroville and the

[23] "The Geology of the Whitestone Mountain Area, Washington" (Master's thesis, Stanford University, 1936).

[24] "On Migmatites and Associated Rocks of Southwestern Finland. Part I," *Comm. Geol. Finland, Bull. 58* (1923), pp. 1–151.

[25] "The Rock Groups of Jersey," *Proc. Geol. Assoc. London*, Vol. XLII (1931), pp. 178–215.

[26] "The Contaminated Granite of Bibette Head, Alderney," *Geol. Mag.*, Vol. LXIX (1932), pp. 433–51.

[27] "Xenoliths of Igneous Origin in the Tregastel-Ploumanac'h Granite," *Quar. Jour. Geol. Soc. London*, Vol. LXXXVIII (1932), pp. 274–96.

[28] "The Oatland (Isle of Man) Igneous Complex," *Proc. Geol. Assoc. London*, Vol. XLIV (1933), pp. 355–77.

[29] "Sur la recrystallization des diorites du plateau de Manson," *Compt. rend. acad. sci.*, Vol. CC (1935), pp. 2091–93.

[30] "The Cairnsmore of Carpshairn Igneous Complex," *Quar. Jour. Geol. Soc. London*, Vol. XCI (1935), pp. 47–76.

Osoyoos granodiorite, a possible association of the Kruger Mountain alkaline rocks with the Similkameen granodiorite, and probable marked age differences among some of the intrusions because of their conspicuous differences in structure. These meager data and speculations show the necessity for a detailed study of the problem of correlation.

The approach to the problem in the discussion below is in part theoretical, in part empirical—the theoretical part based on our present knowledge of the behavior of igneous intrusions, the empirical part on comparisons among the plutonic rocks of the Okanogan Valley.

RÉSUMÉ OF THE LITERATURE

As far as I am aware, the general problem of correlation has not been attacked previously. As a rule, writers faced with specific instances have either applied uncritically whatever criteria were available or dismissed the question as unsolvable. One exception is the recent work of a few English geologists, who have examined very carefully the possibility of correlation by means of heavy minerals.

A. Brammall and H. F. Harwood[31] cite a long list of criteria used in differentiating the two chief granitic types at Dartmoor, England, but do not discuss their validity. The list includes topographic and structural relationships, textures, heavy-mineral content, bulk composition, kinds of inclusions, amounts of feldspar phenocrysts, amounts and habits of garnet, fluorite, cordierite, etc., presence of foliation, numbers of dikes and veins. In a later study of detrital minerals from the Dartmoor granites Brammall[32] concludes that relative amounts of individual minerals give little information about their origin but that the composition of an entire assemblage, peculiar varieties of individual minerals, and characteristic associations give at least a basis for intelligent guessing. More recently Brammall and S. Bracewell[33] point out the fallacy of basing a correlation on special properties of garnet, cordierite, or mica, since these three

[31] "The Dartmoor Granite: Its Mineralogy, Structure, and Petrology," *Min. Mag.*, Vol. XX (1923), pp. 37–53.

[32] "Dartmoor Detritals," *Proc. Geol. Assoc. London*, Vol. XXXIX (1928), pp. 27–48.

[33] "Variability of Garnet and Cordierite in Granite," *Min. Mag.*, Vol. XXIV (1936), pp. 254–64.

vary considerably over short distances within the same mass. A. W. Groves[34] believes it possible to distinguish between different granites of the Channel Islands and to correlate them with granites of Brittany and southern England by comparisons of heavy-mineral suites; Groves emphasizes the importance of using such early crystallizing primary minerals as zircon and apatite rather than minerals of pneumatolytic or secondary origin (tourmaline, epidote) or minerals introduced by assimilation (andalusite, garnet). Wells[35] and Wells and Wooldridge,[36] who have also worked on granites of this region, do not share Groves's enthusiasm for heavy-mineral correlation but regard presence or absence of foliation as a more satisfactory criterion. Taylor and Gamba[37] approach the problem of correlation with extreme caution; although a granite of the Isle of Man differs from adjacent intrusives considerably in both its essential and its accessory minerals, they hesitate to call it a separate mass. Two years later Taylor[38] concludes from an extensive study of English and Scotch granites that correlation by heavy minerals is in general invalid, because the abundance and kind of minerals are too changeable within the same mass. Waters[39] differentiates the igneous portion of the Swakane gneiss from the adjacent Chelan batholith, near Entiat, Washington, because it contains no zoned plagioclase and no augite and because it has cataclastic structure. G. Baker[40] finds marked differences between several adjacent Victorian granites in heavy-mineral content, in abundance of biotite, in the presence and abundance of hornblende, allanite, calcite, augite, diopside, and

[34] "The Heavy Minerals of the Plutonic Rocks of the Channel Islands," *Geol. Mag.*, Vol. LXIV (1927), pp. 241–51, 457–73; "Heavy Mineral Suites and Correlation of the Granites of Northern Brittany, the Channel Islands, and the Cotentin," *Geol. Mag.*, Vol. LXVII (1930), pp. 218–40.

[35] "The Heavy Mineral Correlation of Intrusive Igneous Rocks," *Geol. Mag.*, Vol. LXVIII (1931), pp. 255–62.

[36] *Op. cit.* [37] *Op. cit.*

[38] "The Correlation of Intrusive Rocks by Means of Heavy Minerals," *British Assoc. Adv. Sci. Rept. 1935* (1935), p. 381.

[39] "The Swakane Gneiss, Entiat Mountains, Washington," *Jour. Geol.*, Vol. XXXV (1932), pp. 158–70.

[40] "Petrology of the You Yangs Granite," *Proc. Roy. Soc. Victoria*, Vol. XLVIII (1936), pp. 124–58.

rutile; he suggests that the differences can be explained by different degrees of contamination in the same granite but makes no attempt to settle the matter conclusively. In discussing the correlation of several isolated masses with the "older" and "younger" granites of Finland, H. Väyrynen[41] uses as criteria the texture of siliceous and basic varieties, the presence of orthoclase, garnet, and two kinds of pyroxene, and the presence and composition of pegmatite dikes; in one case the correlation is made solely on the presence of microcline. M. P. Billings[42] distinguishes the various granites of his area in New Hampshire by dike relations wherever possible, otherwise by mineral composition, amount of shear, and amount of foliation. S. A. Tyler[43] believes that peculiarities of zircon afford a basis for correlation among intrusive rocks in New Jersey and southern New York.

DISCUSSION OF CRITERIA

In general, the best criteria for correlating intrusive rocks are (1) cross-cutting relationships, either of the intrusives themselves or of their satellite dikes, and (2) exact age determinations from either stratigraphic evidence or measurements of radioactive disintegration. Patently, such criteria, when available, must always take precedence over the criteria to be discussed below.

One possible criterion will be dismissed forthwith—the criterion of simple proximity. Evidently rocks from outcrops separated by only a few feet of alluvium are more probably related than those from outcrops separated by several miles; but the application of this truism in specific cases must be largely subjective, depending on the observer's estimate of the variability of the intrusive in question. In general, validity of the criteria to be mentioned becomes highly questionable where outcrops are lacking for more than a few miles.

Difficulties in correlation are especially great for heterogeneous masses like some of the Okanogan intrusions. The detailed analysis of criteria to follow is an attempt to find methods of correlation ap-

[41] "Über die Altersverhältnisse der Granite von Südfinnland und Pohjanmaa," *Comm. Geol. Finland, Bull. 115* (1936), pp. 251–65.

[42] "Regional Metamorphism of the Littleton-Moosilauke Area, New Hampshire," *Bull. Geol. Soc. Amer.*, Vol. XLVIII (1937), pp. 463–566.

[43] "Zircon Studies in the New Jersey Highlands," *Amer. Jour. Sci.*, Vol. CCXXXVIII (1940), pp. 260–71.

plicable to just such variable masses. Obviously, in dealing with more homogeneous intrusives the examination of criteria need not be so scrupulous.

The criteria will be considered under five heads: (1) chemical composition; (2) petrographic characteristics, including texture, mineral composition, and mineral peculiarities; (3) structure, including foliation, cataclasis, and larger features like banding and xenoliths (4) metamorphism, both of the intrusive and of surrounding rocks; and (5) abundance, petrography, and structure of dikes and veins.

Chemical composition.—The usefulness of chemical data in correlation depends in large measure on the number of analyses available. With sufficient analyses to construct reliable variation diagrams, trends of differentiation and assimilation can be established for each mass, so that significant chemical variations between different intrusives will not be confused with variations of the individuals. But comparisons based on the small number of analyses usually available may be grossly misleading.

The simplest chemical distinction between two intrusives would be based on significant and constant differences in the percentages of one or more constituents. Commonly, however, differentiation and assimilation make the percentage composition of each intrusive so variable that this criterion can be safely applied only to such obvious distinctions as that between granite and gabbro. Percentages of some of the scarcer constituents (MnO, BaO, SrO, ZrO_2, P_2O_5) might be more useful, but too little is known of the role of these oxides in the processes of differentiation to put their usefulness beyond question.

Since differentiation, with or without assimilation, commonly leads to the formation of a progressively more siliceous magma with an increase in SiO_2, Na_2O, K_2O, and a decrease in FeO, CaO, and MgO, initial differences between two magmas might be expected to show up more clearly as differences of ratios within these groups (e.g., Na_2O:K_2O, CaO:MgO) than as differences in the individual oxides. But even these ratios are hardly reliable unless considerably and consistently different.

Available analyses of the Okanogan intrusives (in part from Daly's memoir, in part from analyses made for Dr. A. C. Waters) are assembled in Table 1. Analyses I, II, and III should give a

TABLE 1

CHEMICAL ANALYSES

	I	II	III	IV	V	VI
SiO_2	67.77	72.46	68.65	57.80	55.26	64.76
Al_2O_3	17.69	15.04	15.34	17.46	18.58	16.98
Fe_2O_3	*T*	0.57	0.87	1.33	1.41	1.71
FeO	2.79	1.15	2.49	5.94	5.15	2.18
TiO_2	0.48	0.24	0.38	0.88	0.87	0.44
MnO	0.13	*T*	0.07	0.11	0.06	0.15
CaO	3.16	2.71	3.63	7.96	7.69	4.86
MgO	0.58	0.39	1.34	3.87	4.48	1.04
K_2O	1.98	2.13	2.24	0.67	1.32	2.73
Na_2O	4.38	4.70	4.44	2.86	3.83	3.58
H_2O-	0.10	0.19	0.18	0.15	0.05	0.08
H_2O+	0.62	0.30	0.20	0.67	0.78	0.74
CO_2	*nil*	*nil*	*nil*	*nil*	*nil*	0.39
P_2O_5	0.16	0.09	0.22	0.22	0.44	0.23
BaO	*ND*	0.07	0.08	*ND*	*ND*	*ND*
SrO	*ND*	*T*	0.04	*ND*	*ND*	*ND*
Totals	99.84	100.04	100.17	99.92	99.94	99.87
	Norms					
Quartz	25.4	30.2	23.5	13.8	4.0	22.0
Orthoclase	11.7	12.8	13.3	3.9	7.8	16.1
Albite	37.1	39.8	37.7	24.1	32.5	31.0
Anorthite	14.5	12.8	15.3	32.8	29.5	20.0
Corundum	3.1	0.2	.....	.....	.....	0.7
Dps $CaSiO_3$	.....	.....	0.7	2.2	2.4	.....
Dps $FeSiO_3$	.....	.....	0.4	1.1	0.9	.....
Dps $MgSiO_3$	.....	.....	0.3	1.1	1.4	.....
Hyp $FeSiO_3$	4.6	1.2	3.1	7.5	6.1	2.0
Hyp $MgSiO_3$	1.5	1.0	3.0	8.6	9.8	2.6
Magnetite	.....	0.9	1.1	1.9	2.1	2.5
Ilmenite	0.9	0.5	0.7	1.7	1.7	0.9
Apatite	0.4	0.2	0.5	0.5	1.0	0.7
H_2O, BaO, etc.	0.7	0.6	0.5	0.8	0.8	1.7
Totals	99.9	100.2	100.1	100.0	100.0	100.2

I. Colville batholith, homophanous facies (core of Mount Annie)

II. Colville batholith, swirled gneiss facies; average of 21 samples from just south of mapped area

III. Colville batholith, banded gneiss facies; average of 25 samples from just south of mapped area

IV. Colville batholith, banded diorite gneiss; basic layer in banded gneiss facies, 1 mile north of Chewiliken Valley

V. Colville batholith, core of Tunk Mountain

VI. Whisky Mountain stock, average of 4 samples

TABLE 1—*Continued*

	VII	VIII	IX	X	XI	XII
SiO_2	68.43	66.55	66.65	50.49	52.53	55.11
Al_2O_3	15.80	16.41	15.29	15.83	19.05	21.28
Fe_2O_3	1.06	1.98	0.39	6.11	4.77	2.64
FeO	1.85	1.80	3.30	3.04	2.10	1.29
TiO_2	0.20	0.40	0.52	0.92	0.07	0.48
MnO	0.10	0.12	0.12	0.11	0.13	0.08
CaO	4.08	3.86	4.04	7.99	5.75	2.82
MgO	1.46	1.32	1.85	3.38	1.99	0.59
K_2O	2.51	2.84	3.36	6.86	7.30	8.36
Na_2O	3.47	4.07	3.24	3.12	4.03	6.24
H_2O-	0.05	0.01	0.10	0.29	0.13	0.14
H_2O+	0.53	0.24	0.96	1.20	1.49	0.58
CO_2	*ND*	*ND*	0.34	0.07	0.27	0.08
P_2O_5	0.07	0.15	0.19	0.42	0.28	0.27
BaO	0.09	0.03	*ND*	*ND*	0.09	*ND*
SrO	0.02	0.01	*ND*	*ND*	0.19	*ND*
Totals	99.72	99.59	100.35	99.83	100.17	99.96
	Norms					
Quartz	26.8	21.8	22.6	.....	.....	.....
Orthoclase	15.0	16.7	20.0	40.6	43.4	49.5
Albite	29.3	34.6	27.3	7.3	11.0	13.6
Anorthite	19.7	17.5	16.7	8.9	12.0	5.3
Nepheline	.....	.....	.....	10.2	12.5	21.3
Corundum	.....	.....	0.2	.....	.....	.....
Diopside	0.2	0.6	.....	18.1	10.8	3.2
Hypersthene	5.6	4.1	9.7	.....	.....	.....
Wollastonite	.....	.....	.....	2.0	0.7	1.0
Magnetite	1.6	3.0	0.5	7.4	6.7	3.0
Ilmenite	0.5	0.8	1.0	1.7	0.2	0.9
Hematite	.....	.....	.....	1.0	0.2	0.5
Apatite	0.3	0.3	0.4	0.9	0.6	0.6
Calcite	.....	.....	0.8	.....	.....	.....
H_2O, BaO, etc.	0.7	0.3	1.0	1.5	1.6	0.7
Totals	99.7	99.7	100.2	99.6	99.7	99.6

VII. Osoyoos batholith, "original granodiorite," from point 2 miles north of Canadian border and 2 miles west of eastern contact

VIII. Similkameen batholith, normal type, from west side of Similkameen Valley 3 miles north of Canadian border

IX. Similkameen batholith, normal type, from point 2.5 miles east of Lemanasky Lake, Chopaka quadrangle

X. Malignite from Kruger Mountain

XI. Femic nepheline syenite from Kruger Mountain

XII. Nepheline syenite from Kruger Mountain

Analyses I–VI, and IX by F. Herdsman, Glasgow; analyses made for Dr. A. C. Waters, and used with his kind permission. Analyses VII, VIII, X–XII copied from Daly (pp. 440, 450–53, 456).

ND = not determined; *nil* = not found; *T* = trace.

reasonably accurate idea of the average compositions of the three zones of the Colville batholith; IV and V represent basic portions of this batholith. Analysis VI is a composite sample from the Whisky Mountain granodiorite and VII a single sample representing the least metamorphosed variety of Osoyoos granodiorite. Analyses VIII and IX represent two widely separated samples from the Similkameen batholith. Analyses X, XI, and XII show the compositions of the three principal varieties among the alkaline rocks of

TABLE 2

MOLAL RATIOS

(All Fe Calculated as FeO)

Analysis	Intrusive	MgO:FeO	CaO:MgO	FeO:CaO	K_2O:Na_2O
I	Colville	0.56	3.9	0.67	0.30
II	Colville	0.42	5.0	0.48	0.30
III	Colville	0.73	1.9	0.71	0.33
IV	Colville	0.98	1.6	0.70	0.15
V	Colville	1.26	1.2	0.65	0.23
VI	Whisky Mountain	0.50	3.4	0.60	0.50
VII	Osoyoos	0.93	2.0	0.53	0.48
VIII	Similkameen	0.66	2.1	0.73	0.46
IX	Similkameen	0.91	1.5	0.71	0.68
X	Kruger Mountain	0.71	1.7	0.84	1.45
XI	Kruger Mountain	0.55	2.1	0.87	1.19
XII	Kruger Mountain	0.29	3.4	1.02	0.88

Kruger Mountain. No analyses are available for the quartz diorite, the Earl Mountain metagabbro, or the Oroville malignite.

The inadequacy of chemical criteria is well illustrated by these analyses. The Kruger Mountain alkaline rocks are, of course, readily distinguished from the others by their low silica and abundant alkalies, but among the granitic rocks nearly every oxide or combination of oxides shows a greater variability within the Colville or Similkameen batholiths than between different masses. Ignoring the two basic Colville samples helps little. Possibly the low P_2O_5 content of the Osoyoos sample may be significant, but conclusions based on a single analysis are scarcely safe.

Ratios of various oxides are listed in Table 2. The only significant fact which these bring to light is the consistently low K_2O : Na_2O ratio in the Colville rocks.

Petrographic characteristics.—Mineral composition as a criterion for correlation is subject to the same limitations which differentiation and assimilation impose on chemical composition but has the important advantages of a larger number of variables and much easier determination.

A mineralogical criterion often used uncritically in correlating both sedimentary and igneous rocks is the abundance and kinds of the less common minerals present. The validity of such data has been studied in detail by the English writers mentioned on pages 32 and 33, particularly by Brammall and Taylor. Their conclusions are in general pessimistic, since even the rare minerals vary unpredictably within a single intrusive. From their work come several rules for the application of this criterion to intrusive igneous rocks: (1) each intrusive must be thoroughly sampled; (2) assemblages of rare minerals give better evidence than individuals; (3) peculiar characteristics of a mineral constant throughout an intrusive are much more valuable than its mere presence or abundance; (4) minerals indigenous to an intrusive (e.g., apatite, zircon, allanite in granite) are safer than those which may have been introduced by assimilation or metamorphism (e.g., sphene, garnet, magnetite). Even in the best of circumstances a correlation based on rare minerals alone is not altogether safe.

Among the Okanogan rocks *zircon* is fairly abundant in the Osoyoos and Colville granodiorites, rare in the Similkameen granodiorite, absent or rare in the other intrusions. More significant is its characteristic occurrence as tiny, very slender rods in the Colville granodiorite. Although several Colville specimens do not show these odd zircons, they are apparently restricted to this batholith. *Allanite* in small amounts is a common, but by no means universally present, accessory in Colville rocks, its amount showing no relation to position in the batholith. The mineral was not observed in the other intrusions, although Daly reports it from the "basified contact zone" of the Osoyoos batholith. *Apatite* is considerably less abundant and in smaller crystals in the Osoyoos batholith than in the other intrusions (probably reflected in the low figure for P_2O_5 in Analysis VII). *Sphene* is nearly absent in the Colville granodiorite, except in the more basic varieties (which have undoubtedly

been contaminated by migmatization and assimilation) but is a fairly constant accessory in the other masses. In the Whisky Mountain rocks its crystals are often markedly elongated; in the Oroville malignite and in the Kruger Mountain alkaline rocks it commonly occurs as tiny granules surrounding magnetite grains. *Melanite* is a common accessory in the Kruger Mountain alkaline rocks and in syenite associated with the Oroville malignite. Other accessories show no significant differences among the intrusives. To summarize: The slender zircons, the presence of allanite, and the absence of sphene are fair evidence that the Colville granodiorite is distinct from the other intrusions; the presence of brown garnet and of magnetite rimmed with sphene is fair evidence for a relationship between the Oroville malignite and the Kruger Mountain alkaline rocks; no further conclusions seem justified.

Amounts of the more abundant minerals are, of course, the basis for such elementary distinctions as that between gabbro and nepheline syenite or between malignite and granodiorite but are not very useful for finer distinctions among masses of variable composition—a conclusion to which the variegated gneiss of the Colville batholith bears eloquent testimony. The scarcity of epidote and amphibole in the Colville rocks, the absence of amphibole and the low-quartz content in the granodiorite of Whisky Mountain may have some significance; the nearly uniform presence of pyroxene in the Oroville malignite and in the Kruger Mountain rocks helps to distinguish them from the Earl Mountain metagabbro (see Table 3). But this criterion is entirely safe only in regions where intrusives are more homogeneous.

Abundance ratios between certain minerals are sometimes informative. In the normal course of differentiation toward a more siliceous magma, potash feldspar and quartz should increase approximately together while plagioclase decreases (and becomes more sodic). Likewise, in the more common assimilation processes potash feldspar and quartz would be expected to increase or to decrease approximately together. Thus, two outcrops which differ considerably in the ratio (plagioclase): (potash feldspar + quartz) might easily belong to the same intrusion; but, if the ratio (potash feldspar: quartz) is markedly different, the relationship would be dubi-

ous. For instance, a large amount of potash feldspar with very little quartz at once distinguishes the Oroville malignite and the Kruger Mountain alkaline rocks from the other intrusives. Among the granitic rocks, the low (potash feldspar: quartz) ratio in the Osoyoos

TABLE 3

ROSIWAL ANALYSES

	A	B	C	D	E	F	G	H
Quartz	31%	34%	18%	11%	19%	37%	20%	22%
K-fs*	7	22	.....	2	22	8	32	34
Plag	50	35	43	48	48	33	41	30
Biotite	11	7	10	16	6	11	6	5
Hb	.....	.....	27	21	.....	3	1	4
Magn	.....	1	2	2	1	.....	.....	2
Sphene	.....	.....	.....	.....	1	.....	.....	1
Epidote	.....	.....	.....	.....	3	8	.....	1
Apatite	.....	.....	.....	.....	.....	.....	.....	1
Musc	1	1	.....	.....	.....	.....	.....	.....

A. Colville batholith, homophanous facies (core of Mount Annie, 10 miles southeast of the southeast corner of the mapped area)

B. Colville batholith, homophanous facies (Mount Hull)

C. Colville batholith, basic layer in banded gneiss (8 miles south of south border of mapped area)

D. Colville batholith, basic type (core of Tunk Mountain, 13 miles south of southeast corner of mapped area)

E. Granodiorite of Whisky Mountain (phenocrysts included by combining microscopic analysis of groundmass with megascopic analysis of hand specimens)

F. Osoyoos batholith, "original granodiorite" (from a point two miles north of the border and two miles from the eastern contact); taken from Daly (p. 440)

G. Similkameen batholith, normal type (Toat's Coulee, two miles west of west border of mapped area)

H. Similkameen batholith, normal type (west side of Similkameen Valley, three miles north of border) taken from Daly (p. 456)

* K-feldspar includes orthoclase, microcline, and microperthite.

granodiorite gives a possible reason for associating it with the quartz diorite north of Oroville and for distinguishing it from the other siliceous intrusives; a fairly high (potash feldspar: quartz) ratio in the Similkameen and Whisky Mountain granodiorites suggests a possible relationship between these masses.

Abundance ratios among the ferromagnesian minerals are highly variable in normal assimilation and differentiation processes, hence

can only rarely be satisfactory for correlation. Both the Osoyoos and Colville granodiorites bear witness to the extreme biotite: amphibole variation possible within a single intrusive, and the Kruger Mountain alkaline rocks show a similar inconstancy in the (amphibole : pyroxene) ratio.

Properties of individual minerals should presumably provide good clues for correlation, although their usefulness is severely limited by our ignorance of many of the factors responsible for the properties. The mode of occurrence of *quartz* may sometimes be helpful in characterizing an intrusive—for instance, pockets of flamboyant quartz, probably a consequence of crushing and the movement of silica-rich solutions, are peculiar to the Osoyoos and Colville granodiorites. Fluid inclusions in quartz should sometimes give valuable information for correlation,[44] since the composition of the fluid should depend on conditions of temperature, pressure, and composition in the original magma; some doubt is cast on the usefulness of this criterion by W. H. Newhouse,[45] who points out the difficulty of distinguishing primary and secondary inclusions. Fluid inclusions in the Okanogan rocks were not large enough or abundant enough for detailed study. *Potash feldspar* may differ from one intrusion to another in the size of its crystals, in its tendency to fill irregular spaces between other crystals, in the presence or absence of microcline twinning, and in the abundance of perthitic intergrowths; but, since none of these are demonstrably connected with differences more radical than slight temperature and pressure variations during solidification, they have little correlative value. Among the Okanogan intrusives, potash feldspar in the Oroville malignite, in the alkaline rocks of Kruger Mountain, in the Whisky Mountain granodiorite, and in the Similkameen granodiorite is mostly microcline and microperthite; in the quartz diorite north of Oroville and in the Osoyoos batholith mostly orthoclase; in the Colville granodiorite and the Earl Mountain metagabbro sometimes microcline and sometimes orthoclase. Phenocrystic development is almost universal in the

[44] W. Mackie, "The Sands and Sandstones of Eastern Moray," *Trans. Geol. Soc. Edinburgh*, Vol. VII (1897), pp. 148–72.

[45] "The Composition of Vein Solutions as Shown by Liquid Inclusions in Minerals," *Econ. Geol.*, Vol. XXVII (1932), pp. 419–36.

Whisky Mountain intrusion, common in parts of the Colville and Similkameen batholiths. But only in the Earl Mountain metagabbro do the veinlets and replacements of potash feldspar give it a distinct value in correlation.

The composition of *plagioclase* changes too readily with differentiation and assimilation to be very useful in comparing different intrusives. Some weight can be attached to it, however, when it remains fairly constant over wide areas and when it is markedly different from one intrusive to another. Thus, the common plagioclase of the Osoyoos batholith (An_{38}) is conspicuously more calcic than the average plagioclase of any other siliceous intrusive in this region; on the other hand, average plagioclases in the Whisky Mountain (An_{25}), Colville (An_{25}), and Similkameen (An_{30}) rocks are not significantly different in composition. The presence and kind of plagioclase zoning should be of particular importance in correlation, since zoning reflects a range of conditions which covered a long part of the cooling history of an intrusion and which, therefore, should be widespread rather than local. Just how similar zoning should remain in different parts of an intrusive cannot be stated precisely, but some variation seems reasonable. Thus, the constant presence of zoned plagioclase (mostly normal but in part oscillatory) in the Whisky Mountain and Similkameen intrusions shows a probable relationship, despite the more marked zoning in the latter. Development of clear albitic rims on plagioclase should be infrequent enough to have some diagnostic value; such rims appear in the Whisky Mountain granodiorite, the Kruger Mountain alkaline rocks, the Oroville malignite, and to a less extent in the Similkameen granodiorite.

The correlative importance of myrmekite is a matter for conjecture, since conditions necessary for its formation are not clearly understood. If Na-bearing solutions in a late magmatic stage are responsible, it should have at least as great a value as albitic rims on plagioclase, particularly if widespread through an intrusion. In this region myrmekite is fairly abundant throughout the Osoyoos and Colville rocks, rare in the Whisky Mountain and Similkameen intrusion, and absent elsewhere.

Since *biotite* may originate in several ways—by direct crystalliza-

tion, in late magmatic stages from amphibole or pyroxene, or by later metamorphism—and since its composition and properties are widely variable, it should be a valuable mineral for purposes of correlation; but, unfortunately, the relations between the properties of biotite and its history are yet too imperfectly known. The aggregates of tiny green crystals in old amphibole shapes which characterize the Whisky Mountain granodiorite set this intrusion off sharply from the others, but no further significant differences in the various biotite varieties were observed.

Amphibole and *pyroxene* likewise should be sufficiently responsive to the conditions affecting a particular intrusion to be useful in correlation. The distinctive hastingsite and unusual augite of the Oroville malignite, for instance, associate it clearly with the Kruger Mountain alkaline rocks. The amphibole of the Osoyoos granodiorite, although not as definitive as hastingsite, helps to distinguish it from the Similkameen granodiorite and from the amphibole rocks of the Colville batholith.

The value of *epidote* in correlation remains problematical, since the possibility of its primary origin is still debated. Conceivably, marked differences in the composition of epidote between two intrusions might serve to distinguish them, but no such differences are found in this region. Differences in its form are conspicuous: in the Osoyoos batholith it occurs as large clear crystals associated with biotite and hornblende, pseudomorphing neither, apparently earlier than some of the biotite, to all appearances a primary mineral; in other intrusions it is for the most part clearly secondary.

Unusual *textures*, like the famous "rapakivi texture" of a Finnish granite, may have considerable correlative value. Reaction rims, except when produced by local assimilation, are a textural feature of great importance in characterizing an intrusion, since they reflect peculiar conditions of cooling during an interval in a rock's cooling history. Such ordinary textures as occur in this region (exclusive of cataclastic and foliated textures, which will be considered in the next section), however, vary so greatly within a single mass that their value for correlation is slight. Normal granitic textures, modified somewhat by cataclasis, foliation, and material introduced from solution, are the commonest type. In the Osoyoos and Colville gran-

odiorites occasional areas show allotriomorphic textures. A porphyritic texture is nearly constant over the Whisky Mountain intrusion and appears in still larger areas within the Colville and Similkameen batholiths. The only textural features with probable significance in correlation are the amphibole-shaped biotite-epidote aggregates in the Whisky Mountain granodiorite, the platy biotite-hornblende-epidote aggregates of the Osoyoos granodiorite, and the hastingsite-augite association in the Oroville malignite and the Kruger Mountain intrusive.

Structure.—Large features like xenoliths, schlieren, and irregular basic banding, which can be reasonably interpreted as a result of incomplete assimilation of foreign material, should provide one of the safest means of correlation; for conditions of temperature, pressure, and composition favorable or unfavorable to the assimilation of large amounts of foreign matter should, in general, be characteristic of an intrusion as a whole rather than of a small part. True schlieren and angular xenoliths are rare in the Okanogan intrusions: they are completely absent in the Whisky Mountain intrusion, present in small numbers along the Osoyoos border, and sometimes outlined in the banded gneiss zone of the Colville batholith. Their presence or absence in the Similkameen batholith has not been commented on (except for the small areas of intrusive breccia at the malignite contact on Kruger Mountain). But the extraordinarily wide zone of banded, probably migmatitic, gneiss in the Colville batholith furnishes one of the best evidences for the distinctness of this intrusion. Some of the basic rocks along the Osoyoos border may be a small-scale imitation of this assimilative behavior but are so much scarcer and so different in appearance that no confusion results.

Gneissic and cataclastic structures may appear only near the margins of an intrusive or, if more widely developed, may show considerable variation within a single mass (the Colville batholith is an extreme example). Nevertheless, major structural features are ordinarily sufficiently continuous to provide a fairly satisfactory means of correlation. The Osoyoos and Colville granodiorites both show well-developed cataclastic and gneissic structures, but the usual absence of linear structure in the Osoyoos rocks is one good reason for

separating the two batholiths. The scant signs of cataclasis and the lack of foliation in the Whisky Mountain granodiorite distinguish it sharply from the adjacent, strongly banded Colville gneiss and suggest its association with the likewise structureless Similkameen granodiorite.

Not only the presence and kinds of structural features but structural patterns also may be important in characterizing an intrusion, since the patterns are determined by late magmatic movements or later deformation of the mass as a whole. For example, the broad, gentle folds in the Colville platy structure contrast with the uniform and moderately steep dips in that of the Osoyoos batholith.

Besides helping to decide questions of similarity between different intrusives, gneissic and cataclastic structures are important in that they furnish the only criterion in this list applicable to age relationships. Their use in chronology depends on the simple proposition that the least deformed of several intrusions should be the youngest—an assumption which has been used without comment by several authors, including Daly for intrusives of this and adjacent regions. Reasonable as the assumption appears, it must be used with considerable caution. In the first place, two intrusives subjected to the same deforming forces can give similar structures only if they are of comparable size and rigidity. If one were not entirely solidified at the time of deformation, for instance, it would presumably yield more readily and despite its younger age would give evidence of more profound structural modification. A second difficulty in applying the assumption is the lack of any scale for measuring degree of deformation. If one mass shows well-developed gneissic structure, another intimate cataclasis, a third mylonitization along certain shear zones, how can it be decided which represents the most profound deformation?

The Okanogan intrusives provide good illustrations of these difficulties. Cross-cutting pegmatite proves that the Earl Mountain metagabbro is older than the Colville granodiorite, yet the former shows no gneissic structure (except in dikes) and much less cataclasis than the adjacent batholith; possibly its smaller size is responsible for the apparent anomaly, possibly its greater rigidity at the time of deformation, possibly later recrystallization. Cataclasis in the Oroville

malignite has produced intense mylonitization along shear zones but little change in the intimate structure of the rock. Superficially, this seems a lesser degree of deformation than that in the adjacent Colville and Osoyoos granodiorites, which suggests a younger age for the malignite. Yet the difference may be due rather to deformation of the granodiorites while in a semiplastic state and of the malignite in a solid state. A similar difficulty is encountered in comparing deformation in the Osoyoos batholith with that in the quartz diorite north of Oroville and in the alkaline rocks of Kruger Mountain.

Metamorphism.—Metamorphism within an intrusion may provide a basis for correlation if sufficiently widespread, but metamorphic changes are frequently more pronounced in one part of a mass than in others. Various metamorphic and late magmatic changes have been discussed under "Petrography" and "Structure."

The kind and extent of exomorphism near an igneous mass give an accurate picture of temperature and pressure conditions at the time of intrusion, and in general should be characteristic of the mass as a whole rather than of a small part. For correlating intrusions, therefore, metamorphic activity should be a reliable criterion.

That application of this criterion may be difficult for a complex intrusive is well illustrated by the Colville batholith. The migmatite zone of this batholith is ample evidence of a once-great metamorphic activity, but along its present margins metamorphic effects are notably lacking. Waters[46] reports that locally south of this area the migmatite zone is thin or absent and contact metamorphism at present levels is much more pronounced. In spite of these complications, the small amount of metamorphism along the present Colville contacts of this area seems good evidence for the distinctness of the intrusion.

Contact metamorphism around the Osoyoos batholith is greater than for any other of the Okanogan intrusives. Greenstones are converted to amphibolites, phyllites to biotite schists and diopside rocks; there is some evidence for assimilation of both types. Similarity in degree of metamorphism is one good reason for regarding the rocks west of Osoyoos Lake as part of the Osoyoos batholith.

[46] Personal communication.

Contacts of the Whisky Mountain and Similkameen intrusions are characterized by mild and uniform baking—another reason for linking these two masses. The Earl Mountain metagabbro locally produces strong contact metamorphism, giving garnet rocks and amphibole rocks. Around the Oroville malignite and the Kruger Mountain intrusion metamorphism is nearly absent.

Dikes and veins.—Although dikes are commonly more plentiful in some parts of an intrusion than in others, their general abundance and compositions should have considerable correlative value. The scarcity of dikes around the Colville batholith, the profusion of aschistic dikes on Whisky Mountain, and the abundance of lamprophyres near the Osoyoos batholith provide good reasons for distinguishing the three intrusions. Both aschistic and diaschistic dikes are reported from the Similkameen border; perhaps it is significant, however, that along those parts of the batholith's margin nearest to Whisky Mountain aschistic types predominate.

Veins also should reflect general conditions of temperature, pressure, and composition in the late magmatic stages of an intrusion, hence should aid in correlation. From the other siliceous intrusives the Colville batholith is set off conspicuously by the scarcity of quartz veins in its vicinity.

SUMMARY

SUMMARY AND CLASSIFICATION OF CRITERIA

The following list is at once a summary of the criteria under discussion and an attempt to group them according to their general applicability.

The general problem of correlation may be stated thus: Given an outcrop or a group of outcrops of an intrusive rock, isolated from an adjacent intrusive of known properties; to find whether the unknown mass is a part of the known intrusive or is a separate igneous body. The solution depends on an enumeration and evaluation of the following criteria—assuming, of course, that exposures are adequate for the study of each criterion used. A criterion of the A group refers to a characteristic of the unknown mass which, if present, links it definitely with the known intrusive and which, if absent, makes a relationship unlikely. A criterion of the B group refers to a characteristic of the unknown mass whose presence strongly suggests re-

lationship or lack of relationship but whose absence gives no evidence one way or the other. A criterion of the C group must be used with caution; it may be useful for some intrusives but not, as a rule, for variable ones. In other words, an A-group characteristic should

TABLE 4

	GROUP
Chemical criteria	
Variation diagrams	A
Percentage of individual oxides	C
Ratios of certain elements	A
Petrographic criteria	
Percentages of individual minerals	C
Ratios of certain minerals	B–C
Peculiar properties of certain minerals	B–C
More common properties	
Fluid inclusions in quartz	B
Nature of K-feldspar	C
Composition of plagioclase	C
Presence and nature of plagioclase zoning	A
Composition and properties of mafic minerals	B
Textures	
Unusual textures (such as rapakivi, orbicular, etc.)	A–B
Porphyritic texture	C
Intergrowths	B–C
Reaction rims	A–B
Associations of minerals	B–C
Structural criteria	
Schlieren and xenoliths	B
Cataclastic structure	B
Gneissic structures	
Presence	B
Kind	A
Pattern	A
Criteria based on metamorphism	
Metamorphism of the intrusive	C
Extent and character of exomorphism	A
Criteria based on dikes and veins	
Presence and number	B
Kind of dikes (aschistic or diaschistic)	B

remain constant throughout a variable intrusive, a B-group characteristic should be constant in the greater part of the intrusive, and a C-group characteristic may be expected to vary erratically. In correlations among more homogeneous intrusives, of course, C-group criteria may often be used with confidence.

Even under favorable circumstances a correlation based on a single criterion can have little value. Only by a careful consideration of several may the probability of a relationship or lack of one be established.

SUMMARY OF CORRELATIONS AMONG THE OKANOGAN INTRUSIVES

Suggested correlations for the Okanogan rocks have become evident in preceding paragraphs. The Oroville malignite is believed to be a part of the Kruger Mountain alkaline intrusion, and the Earl Mountain metagabbro is distinct from both. The Whisky Mountain, Osoyoos, and Colville granodiorites are separate intrusions. The quartz diorite west of Osoyoos Lake is part of the Osoyoos batholith, and the Whisky Mountain granodiorite is related to the Similkameen batholith.

No evidence suggesting relationships between the smaller basic masses and the siliceous intrusives was found. Daly[47] lists five points of resemblance between the Similkameen granodiorite and the Kruger Mountain alkaline rocks which to him seem "family traits suggesting that both belong to one petrogenetic cycle": (1) similar polarization patterns in potash feldspar; (2) rims of olive-green amphibole enclosing pale-green augite (Similkameen contact zone); (3) similar amphibole (this must also refer to contact-zone samples, for the pale-green hornblende from other Similkameen specimens has little resemblance to the dark hastingsite of the alkaline rocks); (4) brown biotite with similar optical properties; (5) similar varieties of augite (Similkameen contact zone). Since three of these resemblances refer to a basic border facies of the Similkameen batholith which, as Daly suggests, may be the result of assimilation of malignite, the correlation seems dubious.

To show how the foregoing list of criteria may be used, the weighing of evidence for the three most troublesome correlations among the Okanogan intrusives is shown in detail below. In each case criteria favorable to the suggested correlation are listed without comment under "Pro," those against it under "Con," in what I believe to be their approximate order of importance. Evidence that the Osoyoos (Os.) and Colville (Col.) granodiorites belong to separate intrusions:

[47] *Op. cit.*, p. 459.

TABLE 5

PRO	GROUP
1. Contact metamorphism more intense near Os.	A
2. Abundant migmatite in Col.	B
3. Slender zircon rods in Col.	B
4. Unusual amphibole in Os.	B
5. Linear structure well developed in Col.	A
6. Dikes more abundant in and near Os.	B
7. Epidote almost universal in Os.	C
8. Amphibole abundant only locally in Col.	C
9. Sphene scarce in Col.	C
10. Allanite a common accessory in Col.	C
11. Small K_2O:Na_2O ratio in Col.	B
12. Apatite scarce, in small crystals in Os.	C
13. Low P_2O_5 in Os.	C
14. Plagioclase more calcic in Os.	C
15. Low K-feldspar in Os.	C
16. Platy aggregates of hornblende, epidote, biotite in Os.	C
17. K-feldspar almost entirely orthoclase in Os.	C
CON	
1. Gneissic and cataclastic structures prominent in both	B
2. Similar orientation of platy structures in adjacent parts	B
3. Fairly abundant myrmekite	C
4. Patches of intricately interlaced quartz	C

Evidence that the Whisky Mountain (Wm.) and Similkameen (Sim.) granodiorites are related:

TABLE 6

PRO	GROUP
1. Extent and nature of content metamorphism similar	A
2. Gneissic and cataclastic structures nearly absent	B
3. Similar zoning in plagioclase	A
4. Clear albitic rims on plagioclase	B
5. Myrmekite scarce	B
6. Aschistic dikes abundant	B
7. Secondary epidote fairly abundant	C
8. Porphyritic texture universal in Wm., in part of Sim.	C
9. Zircon rare except near border of Sim.	C
10. Rather high K-feldspar: quartz ratio	B
11. K-feldspar mostly microcline and microperthite	C
12. Fairly low SiO_2	C
CON	
1. Aggregates of tiny green biotite flakes in Wm.	B
2. Elongated sphene crystals in Wm.	B
3. Considerable hornblende in parts of Sim., none in Wm.	C

Evidence that the Osoyoos granodiorite and the quartz diorite (Qd.) north of Oroville are related:

TABLE 7

PRO	GROUP
1. Similar strong contact metamorphism	A
2. Lamprophyres the most common dikes	A
3. Qd. resembles hydrothermally altered parts of Os.	C
4. Low K-feldspar	C
5. Similar orientation of structures	B
6. Considerable epidote, seemingly as a primary mineral	C
7. K-feldspar almost all orthoclase	C
CON	
1. Little foliation or intimate cataclasis in Qd.	B
2. Normal hornblende in Qd., somewhat anomalous in Os.	B
3. No myrmekite in Qd.	C
4. No pockets of intricately interlaced quartz in Qd.	C
5. No observed zircon in Qd.	C

SUMMARY OF AGE RELATIONS

Three conclusions about the relative ages of the Okanogan intrusions can be based on contact exposures, and a fourth on inclusions: (1) the Earl Mountain metagabbro is older than the Colville granodiorite; (2) the Kruger Mountain alkaline rocks are older than the Similkameen granodiorite; (3) the Colville-Osoyoos contact relations are best explained by assuming that these intrusives are approximately contemporaneous; and (4) the malignite near Oroville is older than the Osoyoos granodiorite.

Further estimates of relative ages must depend on the criterion of degree of deformation. Thus the lack of crushing and foliation in the Similkameen batholith probably implies a younger age than that of the Colville and Osoyoos batholiths. Whether this sort of reasoning can be extended to masses of widely different sizes and compositions is questionable. Its application here would suggest that the Earl Mountain metagabbro is younger than the Colville granodiorite and that the alkaline rocks of Kruger Mountain are younger than the Osoyoos granodiorite. Both of these conclusions are patently false.

Hence the probable chronological sequence of the intrusives is (1) the alkaline rocks and the metagabbro of Earl Mountain (relative age unknown); (2) Colville granodiorite and Osoyoos granodiorite (including the quartz diorite west of Osoyoos Lake); and (3) Similkameen granodiorite (including the Whisky Mountain intrusive).

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FLINT'S FILL HYPOTHESIS OF ORIGIN OF SCABLAND

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ABSTRACT

Flint's fill hypothesis in explanation of the scabland features of eastern Washington, which bases the deposition-excavation sequence upon the rise and decline of glacial Lake Lewis, has the advantage of reducing the events and conditions to terms of streams of moderate size but, nevertheless, has serious shortcomings which make it unacceptable.

INTRODUCTION

The fill hypothesis of R. F. Flint[1] is a welcome addition to the list of proposed explanations of the scabland features of eastern Washington.[2] In basing the deposition-excavation sequence upon the rise and decline of glacial Lake Lewis, it has the apparent advantage of reducing the events and conditions to terms of ordinary streams of moderate size. Nevertheless, it has serious shortcomings, which are considered under the following heads: (1) difficulties involved in eroding multiple channels and in removing the fill in the manner suggested, (2) questionable correlations of scabland deposits, (3) its inapplicability to the peculiar shapes of certain deposits along the drainage routes, and (4) its insufficiency in relation to the general problem and with that the neglect of ice jams, the previous presence of which would form an adequate basis for the resolution of the problems of supernumerary channels, high stream levels, divide crossings, Palouse "islands," reshaped terrace remnants, valley-bottom silts, etc., that seem to constitute such a puzzling array. These objections, too serious to allow acceptance of the fill hypothesis in its present form, and certain advantages of the ice-jam hypothesis are discussed in the following pages. A general review of the whole scabland problem is beyond the scope of this paper.

[1] "Origin of the Cheney-Palouse Scabland Tract, Washington," *Bull. Geol. Soc. Amer.*, Vol. XLIX (1938), pp. 461–523.

[2] A general bibliography is furnished in the work cited.

EROSIONAL PROBLEMS

SCARPS AND TERRACES IN THE "PALOUSE SOIL"

Flint lists a number of multiple scarps eroded in "Palouse soil" that in profile form an irregular sequence not in accord with Bretz's published conception of a "Spokane flood." To bring the streams to the level required to cut such scarps Flint postulates that alluvium filled the adjoining channels approximately to the level of the uppermost scarps. Such thick alluvial fill, however, is not necessary. Instead, if the stream courses became clogged with ice jams, the meltwaters accumulating behind the ice jams would try to escape alongside the obstructions and might readily cut any number of scarps and greatly widen the stream courses in a weak loessial formation. Such scarps might be formed at any level within the range of the waters dammed up and at various times during their rise and decline, even in a very complicated sequence. The greater the number of such scarps, the more irregular their vertical and horizontal distribution, and the wider the system of scabland channels, the more probable is their relation to ice-blocked streams rather than to streams making and later excavating a voluminous fill. The irregular Palouse scarps, though incompatible with Bretz's idea of a Spokane flood, at least do not peculiarly favor the fill hypothesis.

MULTIPLE CHANNELS AND FILL REMOVAL

Although we may agree with Flint that scabland is not absolutely unique but may be found to some degree in other localities, he has dismissed one feature too easily—the multiple channels, including slotlike canyons and notches across spurs. A multiplicity of channels, with many connections, is characteristic of the Cheney-Palouse scabland tract. Flint attempts to explain these peculiar features as the result of "superposition of streams from overlying fill, and by lateral spillover of streams which have aggraded their valleys sufficiently to enable them to flow across former divides." Distributaries across major divides, strikingly illustrated by Devils Canyon and lower Palouse River Canyon, belong in the same category and are so treated by Flint.

The explanation appears plausible until one realizes that during the stage of re-excavation part of the water must flow out Palouse

Canyon, while part of it continues down Washtucna Coulee as "two or more subparallel threads of current"—one to take the Devils Canyon route and the other to continue past Connell—and that these separate courses must maintain their individual integrity until the excavation of the fill is nearly complete.[3] Such continued separation of degrading streams within one valley is contrary to our observations of stream behavior and is especially improbable when the streams working in a single valley are supposed to remove the fill by lateral planation.

The present writer thinks that the maintenance of two or three separate streams, formed by downstream divergences of a larger stream, in Washtucna Coulee during lateral planation of a tract 2–4 miles wide (west of Washtucna) while the surface of the fill is lowered 300 feet or more is impossible. If these few streams seem so extremely unlikely, then what of the many such streams that maintained themselves on the wide tract between Cheney and Washtucna? The mechanism of removal of alluvial fill simultaneously from the various interconnected channels, as advocated by Flint, breaks down utterly. On the other hand, the multiple channels, interconnected courses, and divide crossings are to be expected from a stream system repeatedly clogged by ice jams.

How could several streams, cutting down on gradients between 13 and 20 feet per mile (Flint's figures) meander about on "great areas 10 to 20 miles wide" and virtually complete the task of removing gravel from nearly every route (from all except about 15 per cent of the area) without losing several threads by piracy or by reason of differentials that seem almost inevitable? Would not one or at most a few of the streams gain an advantage, cut off the supply of water from the others, cause abandonment of certain courses, and thereby interrupt removal of the postulated fill at various stages of dissection? A few of the streams, working on such a good gradient, should have incised themselves rapidly in the easily eroded fill. Once the preglacial courses had been recovered after a certain amount of shifting and temporary superposition, such streams could no longer undercut the fill perched on the higher basalt ledges. Terraces pro-

[3] Even then Flint does not explain the basin in Washtucna Coulee "more than 12 miles long and 100 feet deep," which Bretz said no normal stream could erode.

tected by such ledges should be abundant, whereas their present area is small. The fill hypothesis requires a delicate adjustment in the process of downcutting, so that the several streams have time to swing about over tracts up to 20 miles wide, remove most of the fill, and yet not interfere with or intercept one another in the meantime. Even if a thick fill had been built as postulated, it could not have been removed by normal streams in the manner suggested.

Moreover, the removal of such an extensive sheet of thick fill from the Cheney-Palouse scabland tract is probably too great a task for a flow of water "less than that of Snake River today," whether or not the flow was broken up into several courses, unless an inordinate length of time be allowed. By comparison, Yakima, Walla Walla, and Willamette rivers, though generally on gentler gradients and admittedly less vigorous, have removed fill of late glacial age only in strips 1–3 miles wide during all Recent time. Even Columbia River, notwithstanding the flow of water through Grand Coulee during the blockade of the Big Bend by the Okanogan lobe of glacial ice, has not cleared its valley bottom. If the proglacial flow, including the Grand Coulee discharge, had been as effective here as postulated for the Cheney-Palouse area, then the extensive sand and gravel deposit in the Cold Creek–White Bluffs–Richland area, for example (partly overlain by silt equivalent to Flint's Touchet beds), should have been largely removed too, as Lake Lewis would have been in control there also.

Similarly, the floor of Pasco Basin, though concededly flatter, should hardly retain beds of Touchet silt and fine sand within a half-mile of the entrance to Wallula Gateway, where a rock threshold might have predisposed Columbia River to lateral planation. Yet it does retain them! Although most of the fill postulated by Flint is supposed to have been removed from the plateau and from Snake River Canyon and presumably had to pass through the Pasco Basin, where at least an exchange of coarse load for fine would be expected, these fine-grained sediments still remain—not merely near Wallula Gateway but also in and along the lower end of Snake River Canyon itself! It is clear that a voluminous filling (700–800 feet thick and 2–4 miles wide across the top), much of it composed of gravel and sand (to accord with so-called cut-terrace remnants), has not been

taken out of Snake River Canyon since the Touchet beds were deposited. No trace of them is recognizable down the river either. Removal of alluvium from Snake River Canyon has been less than along Columbia River. Then what about the gravel "terrace remnants"? Many of them are true remnants, but they were already remnants when the Touchet beds were deposited. As explained in the next section, their correlation with the Touchet beds is considered erroneous. Thus, an inordinately large dissection of Snake River Canyon fill in post-Touchet time, obviously precluded by the position of the Touchet beds themselves, is obviated.

Suppose for the sake of argument that Snake River Canyon had been filled in late glacial time with alluvium almost as high as the bordering Palouse scarps, while Pasco Basin was filled with Touchet beds little, if any, above the present Touchet scarps and terraces, as in Flint's Figure 6, page 498. The fill surface at Lyons Ferry would have been about 1,325 feet above sea-level; near Snake River Junction about 1,150 feet; near Page about 800 feet; and on the floor of Pasco Basin less than 500 feet above sea-level. That the supply of sand and gravel should have filled the canyon so well, however, and yet not have encroached very much on the lower 15 or 20 miles of the river's course appears very improbable. (Actually the Touchet silts and fine sands lie on a low terrace in the canyon not only at the lower extremity of the river near Burbank, Ash, Levey, and Page but also opposite Snake River Junction, at Walker, at Windust, and near Ayer—this last only a short distance below Lyons Ferry.) But suppose we accept the idea. With the decline of Lake Lewis, whether slow or rapid, Snake River, augmented by proglacial streams from the Cheney-Palouse area, would have had a steep gradient near the Snake River Canyon–Pasco Basin boundary or former lake shore, amounting to 23–32 feet per mile below Snake River Junction. Thus rejuvenated, Snake River should have smoothed out its profile by flushing a considerable quantity of fill from the lower part of its canyon out onto the Touchet beds in Pasco Basin. A post-Touchet fan appears inevitable under such circumstances. Deposited on the basalt ledges south and southwest of Ash siding, for example, it would have had a good chance of preservation. Yet in the area between Ash and Humorist sidings several outcrops between sand

dunes reveal Touchet beds of sand and silt with little coarse sand and only such a small amount of lag pebbles as may have come from the Touchet beds, certainly not the amount which one might expect from a former cover of post-Touchet gravel. As no evidence of a

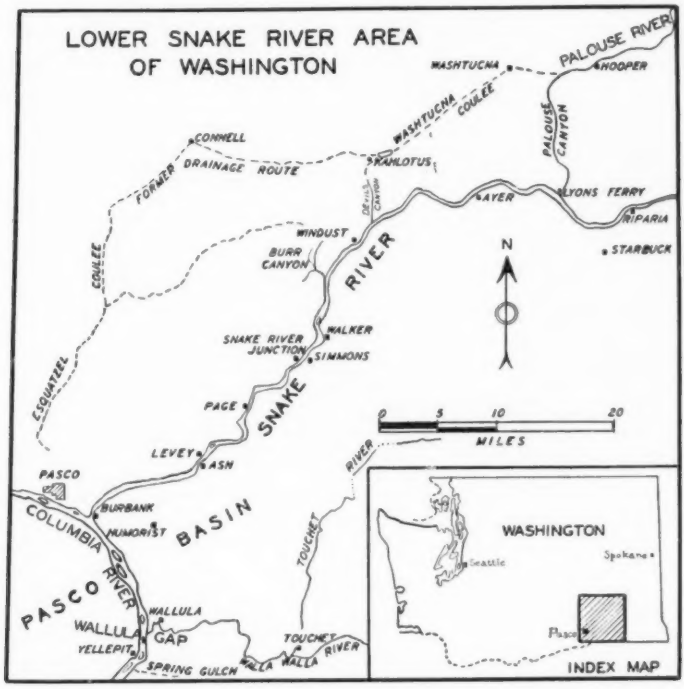

FIG. 1.—Reference map of the lower Snake River area showing the location of places mentioned in the text.

post-Touchet fan is known, Flint's suggested profile of the Snake River Canyon fill in relation to the Touchet beds is subject to grave doubt.

SNAKE RIVER CANYON FILL—TOUCHET BEDS "TRANSITION"

According to Flint, a transition of Touchet beds into scabland fill in Snake River Canyon "is well displayed in cuts on the north

bank, along the Spokane, Portland and Seattle Railway between Pasco and Snake River Junction, and on the south bank, along the Union Pacific Railroad between Ash siding and Walker siding." This generalization is denied here. Typical, fine-grained Touchet beds with characteristic silt dikes occur at railroad levels at Ash, Levey, and Page, but they also extend upriver far beyond Snake River Junction and Walker siding. For example, Touchet beds oc-

FIG. 2.—Terraces along Snake River Canyon in Washington, opposite the mouth of Burr Canyon. The high, even-topped terrace of gravel and sand, nearly filling a former tributary, is a remnant of a former fill of an early stage of alluviation. Below it on the Canyon floor is a low terrace, mainly of sand and silt, not cut from the high terrace as Flint thought but the product of a later and different type of alluviation. Cross-drainage, not abandoned channels of Snake River, modifies the lower terrace.

cur (1) across the river from Burr Canyon, (2) at Windust, (3) near the mouth of Devils Canyon, (4) overlying gravels on both sides of the river near Ayer only a short distance below the mouth of Palouse River (where by Flint's interpretation they should not be), and apparently at other less accessible places seen by the writer only at long range. Throughout this distance the material forms a well-defined terrace 100–150 feet above the river, not merely layers interbedded with gravel and sand (see Figs. 2 and 3). Such a terrace was accepted by Flint but ascribed to a slackening of downcutting during the period of dissection of scabland fill. However, gravel and sand do not occur on top of the silt terrace (except in fans and other

local situations), and no stream channels, except those of tributaries which are extended across it, mar its surface.

Although Flint says a gentle "backward" slope of benches, apparently recording "deeps" at the outsides of bends during fill dissection, is present on many of the low-terrace remnants, the present writer has seen none on the silt terrace at the bottom of the canyon.

FIG. 3.—Terraces along Snake River, opposite Snake River Junction, Washington. The face of the low terrace (darkened with oil spray) appears along the railroad just beyond the river. Behind it stands the high terrace which here has been modified in form so that it appears to be attached to the basalt ledges on the left and to slope off gradually to the valley floor on the right. The sloping appendage, typical of many along the canyon, is thought to have been molded out of the high terrace by waters which swept over its edge while Snake River, then an arm of Lake Lewis, was obstructed by ice-jams. Two gullies have crossed it later. Hills of "Palouse soil" (in the distance) rise abruptly above the high terrace and current-swept benches of basalt at corresponding levels.

Flint's generalization does not apply to it (although applicable at higher levels). Instead, the top of the terrace generally slopes up toward the canyon wall, and locally, as near Page, the silt may be seen to lap well up onto the valley walls, not as slope wash or a wind-blown deposit but as original water-laid beds just as it does in Pasco Basin and in certain valleys in Oregon (similar relations are shown in Fig. 8).

At Levey the terrace top has an undulatory surface to which the bedding as revealed in railroad cuts conforms. Apparently, the sur-

face has been little modified since it was deformed by slumping, minor faulting, and injection of silt dikes. At Windust the railway cuts reveal a rolling surface of strongly foreset cobbly gravel with the overlying beds of silt rising and falling to conform to the uneven gravel surface beneath and thereby making a rolling terrace surface above. These features are understandable on constructional surfaces but not on terraces cut from a thick fill. The persistent occurrence of the silt-capped terrace all along the canyon below Ayer and the lack of river channels or river gravels across it indicate that the terrace surface is one of deposition. Moreover, its profile is such as to join the Touchet beds in Pasco Basin at somewhat less than 500 feet above sea-level without any such sharp change in profile as is required by Flint's correlation. The low-terrace remnants at Lyons Ferry (shown on Flint's Pl. 7) are the upstream equivalent of this terrace. Thus the transition of scabland fill to Touchet beds is much more limited than Flint thought. Let us turn now to the higher-terrace remnants.

Along the same part of Snake River Canyon which displays this silt terrace, Bretz has described a series of gravel and sand deposits that he called "bars." His descriptions of them are reliable. Flint shows a nice, even-crested one in his article (his Fig. 3, Pl. 4). They occur at several places between Lyons Ferry and Levey, at summit elevations ranging from about 1,150 feet above sea-level at the former locality to about 750 feet above sea-level near the latter and locally with thicknesses of 200–400 feet or more. Their summit elevations, where not obviously modified by erosion, seem to conform to a fairly regular profile and to suggest that they are remnants of a single great terrace. The moderately good sorting and rounding shown by the materials suggest that they were deposited under normal conditions, as Flint implies, by an ordinary, overloaded stream and not by the torrential current of the Spokane flood as conceived by Bretz. As the silts on the canyon floor and in Pasco Basin preclude the removal of the remainder of the fill during or since the decline of glacial Lake Lewis, as noted in the preceding section, such removal must have taken place previously. These gravel-terrace remnants are therefore older than Lake Lewis and the silt terrace. This is also in keeping with their high topographic position.

As to age, these high-terrace remnants should be correlated, not with the Touchet beds, but with another terrace (which is comparable in topographic position, bulkiness, grain size, and degree of weathering of its components) well displayed along Columbia River.[4] Near the mouth of Snake River this terrace should originally have been probably between 700 and 750 feet above sea-level. (Flint's estimate of "at least 900 feet" is much too high.) The remnants in Snake River Canyon have about the proper slope to join the above-mentioned Columbia River terrace without any sharp break or knickpoint, such as is demanded by Flint's correlation of these deposits with the Touchet beds.

It may be noted that the gravel bench here referred to lies immediately alongside and about 300 feet or more above some of the best and most accessible exposures of the terrace of Touchet beds on the floor of the canyon, into which, according to Flint, the canyon fill is supposed to grade. The relations of these two deposits may be seen in Figures 2 and 3. Terraces at various intermediate elevations, attributable to cutting from the high terrace, are few. The two deposits are not interbedded except in small areas where the earlier materials were re-worked, as at the following localities.

In Sec. 6, T. 9 N., R. 32 E., near Levey, the gravel is bedded down over the slope and into the gulley, showing that the material has been re-worked since the gulley was formed. Exposures along another gulley in SW. $\frac{1}{4}$ Sec. 6, T. 10 N., R. 33 E., and in Burr Canyon (Fig. 4) show the same thing. In the S. $\frac{1}{2}$ Sec. 5, T. 9 N., R. 32 E., the first railway cut south of the basalt cliff shows basalt rubble, gravel, and sand in beds sloping down over the canyon wall with steep foresets. Apparently, the material was washed off the ledge to the north, some 700 feet above sea-level, over a steep slope into standing water

[4] Although modified by erosion, this terrace occupies considerable parts of the Pasco and Umatilla lowlands, and remnants of it may be found at various other places between Wenatchee, Wash., and Portland, Ore. North of Cold Creek Valley and west of Gable Mountain (Coyote Rapids quadrangle), about 40 miles northwest of the mouth of Snake River, an undissected portion of it stands a little more than 800 feet above sea-level. In the Umatilla lowland, nearly the same distance below the confluence of Snake River with the Columbia, the summit elevation of the terrace is 650–700 feet above sea-level. Downriver its elevation declines to 200–300 feet above sea-level near Portland, Ore.

at least 200 feet deep. The next cut, 300 feet to the southwest, shows a very dirty gravel, sand, and silt mixture resembling concrete aggregate, and the third cut, 450 feet farther on, only fine sand and silt of typical Touchet aspect. The series represents a transition from coarse deposits near the source of supply of such material (the basalt

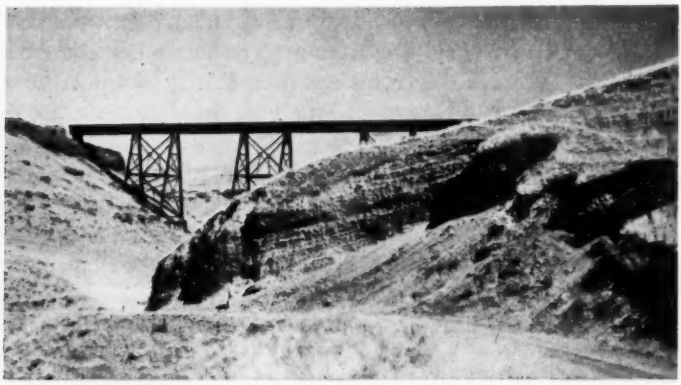

FIG. 4.—Side-hill "bar" in Burr Canyon, a tributary of Snake River. This deposit of silt, sand, gravel, and basaltic rubble was derived mainly from a high-terrace remnant perched on basalt (out of view to the right) after most of the earlier fill equivalent to the high-terrace gravels had been removed from Snake River Canyon (beyond the trestle) and from this tributary. This is not a true bar of a tremendous "Spokane flood" but merely the product of re-working of local materials by waters (necessarily deep but not swift) which swept them off the rock spur and down over the side of the tributary gulch. The beds (essentially long foresets?) dip slightly up tributary but principally across it, opposite to the direction of flow of Snake River. Flint's fill hypothesis has no place for such a feature. Slope wash would not explain its bulging form and probably not the alternations of beds through such a range of particle sizes. Slumping likewise is precluded both by form and structure.

ledge and overlying gravel terrace) through an ill-sorted mixture into fine-grained Touchet beds, all within a distance of only 750 feet.

Another short-range transition is to be seen at Walker, where a large pit in a high gravel terrace (Fig. 5), shows beds deposited during the Touchet stage sloping down to the lower-terrace level on the valley floor. On the north side of Walker Canyon at an elevation of about 550 feet another but smaller pit likewise exposes sand and some silt bedded down the slope in conformity with the present

surface and grading off into Touchet silt beds toward the west and southwest within a few hundred feet. Foresets in the Walker pit, directed down-valley but upslope, as seen in Figure 6, eliminate the possibility of secondary slope wash. The continuity of the beds and the preservation of these foresets likewise eliminate slumping and

FIG. 5.—Terraces along Snake River at Walker siding. The high terrace in the right foreground, composed principally of sand and granule gravel, has a peculiar rounded barlike form. The beds at the outer and upper edges of the pit excavated in the side of the terrace are accordant with the rounded and sloping surfaces. On the up-stream end (near the center) the beds also converge down the slope. The rounded form must have been made by modification of the high terrace after equivalent fill had been removed from most of the canyon. Beds of sand on the gentle slope (to which they conform) in the upper-center grade westward to silt and fine sand which in well-stratified layers make up the low terrace in the left middle distance. The re-working of high-level materials appears to have been simultaneous with the deposition of the materials of the low terrace and bordering slopes.

indicate that the surface on which these beds were deposited was already sloping through a range of more than 250 feet, as Bretz has insisted.

Similarly, in Sec. 25, T. 10 N., R. 32 E., southeast of Page, currents going over the upland swept material toward the southwest and produced a smooth rounded slope not since marred by meander scars or other signs of lateral planation.

The facts recited in this section thus refute Flint's correlation of the gravel terrace in Snake River Canyon with the Touchet beds but indicate a twofold division of the sediments—distinct in texture,

physiographic position, grade profile, and conditions of origin. If Lake Lewis and the Touchet beds are later than much of the gravel, then the rise and decline of Lake Lewis cannot be the cause of the deposition-excavation sequence in Flint's fill hypothesis. Possibly Lake Lewis or equivalent appeared twice, but a special search in the field in 1934 with a two-episode possibility in mind failed to find

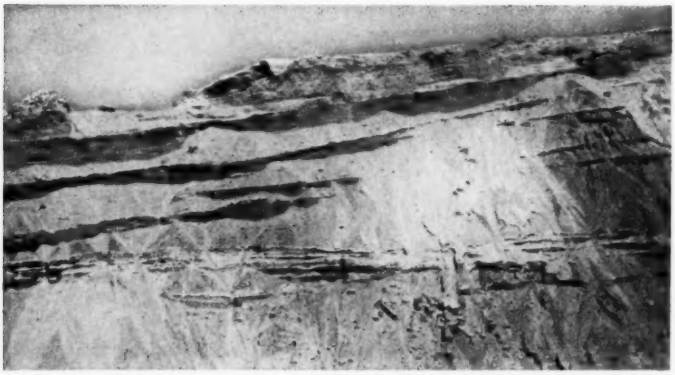

FIG. 6.—Detail of part of the pit at Walker. Note that the foreset bedding is directed up the slope. (Camera was tilted upward somewhat but otherwise carefully leveled from left to right.) Such a structure could not be made by slope wash.

any evidence to support the idea. The continuity of the low-level Touchet beds with the erratic-bearing silts at high levels in Pasco Basin is unmistakable.

CONSTRUCTIONAL FORMS OF SCABLAND DEPOSITS

The consideration of the foregoing features, involving deposition on pre-existing surfaces of considerable relief, leads naturally to our third objection to Flint's fill hypothesis—its failure to account for the peculiar shapes of certain scabland deposits, including the three already mentioned near Levey, Walker, and Page (see Figs. 4, 5, 7, and 8). Bretz described the surface forms of several "bars" along Snake River Canyon and elsewhere. His description of their present shapes is essentially correct, as he recognized that the main canyon and its tributary gulches were not completely filled when these shapes were developed. His mistake was in attributing the "bars" to an enormous flood of high velocity. If the reader will remember

that the gravel and sand were already available along the canyon as high-terrace remnants during the glacial Lake Lewis episode, the origin of the barlike forms becomes comparatively simple. A big flood of high velocity, or a series of such floods, is ruled out by the fine character of the Touchet beds on the canyon floor. Yet the water was not a dead arm of Lake Lewis, or we should have had merely a mantle of silt over the earlier rocks. The runarounds—channels which leave the canyon and rejoin it farther on (as near

FIG. 7.—Barlike terrace beside Snake River, opposite Simmons. This is across the river and downstream from a correlative deposit at Walker (Figs. 5 and 6). The southwest end (at the left) is truncated, but the upstream end is curved and gently sloping. Beyond the terrace are scabland ledges of basalt and a cover of "Palouse soil."

Simmons and near Page[5])—the scabland ledges of basalt scrubbed of all loose material, the sidehill dumps or "shoulder bars," the streamlined forms of the gravel-terrace remnants, and the deposits swept into tributary gulches over their sides definitely indicate currents—in general, currents stronger than those within the canyon. To the writer the only way this combination seems possible is for the canyon to be blocked by ice so that the water of the river was forced to flow along the margins, and thus, impinging against the canyon walls, the currents developed scabland, modified the terrace remnants, cut channels outside the canyon walls where spillovers oc-

[5] These particular ones are exhumed channels of Snake River which antedate the gravel. The superposition of Snake River in its present course likewise preceded the last glacial drainage and is not the product of it as Bretz thought.

curred, cut back Palouse scarps, and otherwise produced an array of features not explainable either by Bretz's Spokane flood or by Flint's fill hypothesis.

Flint attempts to dispose of Bretz's "bars" in part by interpreting them as cut-terrace remnants of a former fill (in which I think he is partly right) and in part by denying any consistent relation between the external form of the deposits and their internal structure (in which, as to re-worked materials, I think he is mistaken). He ex-

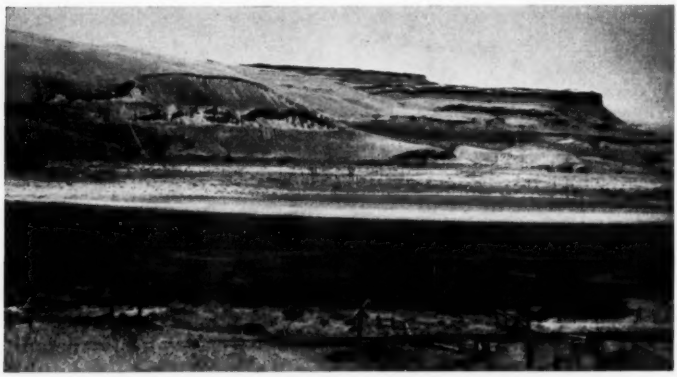

FIG. 8.—Upriver end of the barlike terrace opposite Simmons. Its rounded form and the inclined beds of silt which lie against it are strikingly shown. The altered forms of the high-terrace remnants (but not the terrace itself), the silt beds on the slopes, the low terrace (here eroded), and probably much of the scabland are assignable to the Lake Lewis stage. Hence Flint's fill hypothesis does not apply.

cludes bars. He does not recognize any initial slopes but attributes rounded forms to creep and states that dip-and-strike readings on foreset bedding show little relation to the present surface. His contention is refuted by Figures 4, 5, and 8, which definitely show conformity between shape and structure.

In the case of the Willow Creek "bar" the evidence is perhaps equivocal, although I agree with Bretz that except for minor trenching by Willow Creek the "bar" has a constructional surface, but this does not eliminate the possibility of a general sheet of fill on the Union Flat–Palouse River side of the divide. One may wonder, however, how ordinary streams could have removed the fill and yet have left the numerous "Palouse islands" and the seven channels across the divide west of La Crosse except under conditions of ice blockade.

In their diverse interpretations of the Staircase Rapids "bar" north of Washtucna, both Bretz and Flint are partly right. The flattish top, thin bedding, shortness of foresets, small average grain size, and the occurrence of similar materials with similar structure near Hooper and in Washtucna Coulee farther west support Flint's contention that it is part of a more extensive valley fill; but Bretz's observations of constructional lobings on its eastern side, one of which formerly closed off a part of the gulch, are also upheld. The two are not incompatible, for this seems to be an example of a fill remnant, already terraced when the last glacial waters used the route. The remnant was reshaped by these waters by rounding off the terrace edges and making the crenulations noted by Bretz. Post-glacial erosion in the gulch has been small.

The "bars" at the mouth of Palouse River Canyon do not fit any simple classification. There are several on both sides of Snake River at elevations up to about 1,150 feet above sea-level. These occupy tributary niches or are appended to basalt ledges or knobs and show flutings arranged radially away from Palouse Canyon mouth, as if they had been molded into place. Much of this may be part of a former fill; but its present form, including crescentic scarps, is not the result of ordinary dissection. Instead, it appears that currents out of Palouse Canyon scabland channels disposed the gravel in its present form at a late stage of the glacial drainage by cutting and redepositing materials at hand at several levels above the canyon bottom. A water-ice combination could do this molding without filling Snake River Canyon brimful with gravel. The "midcanyon bar" with its flutings and unfilled depression, already referred to as the upstream equivalent of the low silt-covered terrace, is a different type. It represents the top of the fill made by the last outpouring of glacial waters from Palouse River Canyon. Flint's explanation of it as the product of more or less torrential conditions is correct, but, as noted by Bretz, its top is constructional and not the result of erosion. The flutings (discernible in Fig. 2 of Flint's Pl. 7) are primary features as argued by Bretz. The depression on the terrace or "mid-canyon bar," admitted by Flint to be "unusual," is also primary; it shows no resemblance to deeps at the outside of river bends nor to plunge pools. In my opinion it was left unfilled during the depositional stage because the area was occupied by ice, driven against the

cliff by the force of water from Palouse River Canyon and then protected from wasting by shadow.

The "shoulder bar" in Field's Gulch on the Deruwe ranch is a third type, built by currents which crossed the col in the divide south of Snake River and flowed northwestward down the tributary to rejoin Snake River. On this route Flint and Bretz agree, as well as on the observation that the beds show steep foresets. Bretz considered the deposit a local one; Flint, noting similar gravel across the gulch, took it to be a remnant of a general fill. Here, again, each is partly right. Neither seems to have noticed that the pebbles of these gravels are older than those of the scabland fill in general. Many of the pebbles are cracked and show dull and even pitted surfaces, and some are exfoliating. Ledges of this old gravel crop out farther up the gulch[6] and also on the other side where they evidently were seen by Flint. This high-level terrace of older gravel undoubtedly supplied a large part of the material in the "bar." The material did not have to be carried up out of Snake River Canyon and through the col as Bretz originally thought—hence no catastrophic flow through Palouse Canyon is required. The deposit in its present form is a new one, however, and not a part of the old gravel terrace. It was formerly an important source of gravel for construction purposes, and a large volume was removed; but the high face has been almost entirely obscured by sliding, as only small amounts of gravel have been excavated in recent years. The original form of the deposit can only be inferred; probably it was somewhat rounded into a mound as are other gravel deposits at the ends of distributaries elsewhere. In consideration of the mutilated form of the deposit and of the present conditions of exposure, the strike of the foresets exposed in a 22-foot section, described by Flint, in a deposit "at least 325 feet deep" (according to Bretz) need not be given undue weight in indicating disagreement of form with internal structure. The slope of the rim of the deposit at the upper edge of the pit, however, is at least roughly in the direction of the bedding. The rather steep foresets suggest that the deposit was built into water at the edge of Snake River Canyon by currents which left the major valley a short distance upstream when free flow down the valley was obstructed by ice jams.

[6] The writer first saw these with Dr. Bretz in 1932, but Bretz's papers published earlier do not mention them.

In this section we have attempted to show how the surface features of the scabland deposits as described by Bretz but denied or minimized by Flint and the terrace remnants emphasized by Flint but called "bars" by Bretz can be brought into harmonious and consistent relationships by a proper understanding of the relative ages of the deposits involved. The basis of this reconciliation is the assignment of the high-terrace remnants to an early stage of alluviation followed by an interval of erosion and the assignment of the Touchet beds and the terrace on Snake River Canyon floor to a later episode of glacial drainage obstructed by ice along the drainage routes. Thus obstructed, the waters broke out in divergent channels, crossed divides at many places, cut back Palouse scarps, developed scabland in the underlying basalt, made deposits behind buttes, over tributary walls, and at the mouths of perched distributaries, modified the forms of gravel terraces, etc.

The ice-jam floods and associated Touchet beds are certainly late Wisconsin in age, and the earlier valley filling in Washtucna Coulee, Snake River Canyon, and along Columbia River also belong to some part of the Wisconsin stage. In my opinion the flood waters on the plateau need not have come entirely from glacial ice which had passed over the Spokane River divide, as favored by Flint, but may have been diverted into different routes across the plateau at different times by glacial dams across Columbia and Spokane river valleys at the Big Bend, at Grand Coulee, in the Telford–Crab Creek sector, and near Spokane. Whether the waters so diverted completed the dissection of terraces during a single stage of glaciation before the Lake Lewis–ice-jam stage was reached or whether an interval of ordinary erosion intervened between two distinct glacial stages is uncertain. The first of these alternatives appears more in keeping with Flint's work on the glacial deposits of northeastern Washington, but the occurrence of two Wisconsin substages, perhaps equivalent to the Tahoe and Tioga stages of the Sierra Nevada, would probably fit the regional picture even better.

THE FILL HYPOTHESIS AND THE GENERAL PROBLEM

The assemblage of peculiar physiographic features associated with the scabland, so well publicized by Bretz, requires an explanation consistent with or applicable to all parts of the area from Spokane

and Wenatchee to the mouth of the Columbia River. The clue to one part will probably be found to be the clue to all. Flint's fill hypothesis does not stand this test. In the area between Wallula Gateway and Columbia Gorge, where glacial drainage was more or less ponded to form Lake Lewis, we find the same sort of diversion channels, distributaries, divide crossings, sidehill dumps, local gravel deposits, terrace remnants, valley-bottom silts, scabland ledges, etc., as occur on the plateau of eastern Washington—some of them hundreds of feet above the level of any middle or late Pleistocene alluvial fill. The fill hypothesis is, therefore, of little help here. If ice jams supply a missing key here, probably they were a necessary factor on the plateau also.

Flint himself appears to think that the fill hypothesis should apply to the whole area, for he suggests that the gravels perched at Yellepit and in Spring Gulch on the downstream margins of the Wallula Gateway at elevations up to about 900 feet or more must be the record of an earlier proglacial flow. He also says:

> Wallula Gap also carries well-developed scabland channels up to 1100 feet A.T. (Bretz, 1925, p. 236) which were made by streams presumably when the Pasco Basin was filled with debris up to at least this elevation. The channel may be related to the fill now present, or to a still earlier proglacial fill, or to the Ringold formation. If they date from Ringold time, they have probably been refreshed by flow during one of the glacial stages, because they have a fresher appearance than the channels across the crest of Gable Mountain.[7]

Yet no deposits are known at higher levels in Pasco Basin or in the stream valleys which enter it with which these deposits can be correlated except those associated directly with Lake Lewis and the Wisconsin scabland. Nor is there any evidence that late Pleistocene uplift affected the local area. Even the Ringold formation hardly reaches high enough, and it is far too old. If these deposits are remnants of a formerly continuous fill, the most likely correlatives, as Flint suspects, are those in the Umatilla Basin and in the Pasco–White Bluffs–Cold Creek triangle, but the broad, apparently original top of this last terrace at its highest points west of Gable Mountain and south of Richmond Ferry, about 50 miles above Wallula Gateway, is only 800 feet above sea-level. This fill, therefore, fails to meet the level needed to explain the Yellepit–Spring Gulch deposits

[7] *Op. cit.*, p. 517.

as terrace remnants, much less to explain the fresh scabland at elevations of 1,100 feet or more at Wallula Gap.

The difficulty is easily removed, however, if we admit the possibility that the flow of water through the gap during the high-water stage of Lake Lewis was impeded by ice jams so that currents were diverted through scabland channels over the rim of the gap and along the edges of the ice. After one has seen the coarse (ice-rafted) particles scattered through the Touchet beds, the erratic boulders almost coextensive in distribution with the floods of glacial waters, and the dozens of piles of iceberg-rafted till on the northeast flank of Rattlesnake Hills and in Cold Creek Valley, an ice blockade of the Wallula Gap seems entirely feasible. Icebergs of glacial ice must have been abundant, river ice possibly more so. The overflowing waters could have made local deposits, especially at the lower ends of distributaries, as much as 350 feet above whatever earlier terrace remnants may have been present. If the fill hypothesis, without consideration of ice jams, is inadequate here and at other places along lower Columbia River, then one may properly have misgivings also about its adequacy as the sole explanation of the features in the Cheney-Palouse tract. Drumheller Channels constitute an even stiffer test. If ice was a factor in part of the area, then ice blockades instead of thick alluvial fill may have been instrumental in raising stream levels to unusual heights in Washtucna Coulee, in Palouse River Valley, and elsewhere and thereby to cause the spillovers, distributaries, etc.

CONCLUSIONS

These considerations lead us to conclude that Flint's fill hypothesis (1) imposes a task on the eroding streams which requires decidedly abnormal, if not impossible, behavior of the several strands of current; (2) involves a misinterpretation of the ages of certain gravel deposits relative to that of the Touchet beds; (3) fails to consider adequately the rounded and sloping forms of some of the gravel deposits; and (4), as it is not applicable to Wallula Gap and other scabland areas, is not a general solution of the scabland problem. On the other hand, none of these objections is prejudicial to the idea that these scabland features were developed under conditions of glacial drainage clogged by ice jams.

SUBRIVER SOLUTION CAVITIES IN THE TENNESSEE VALLEY

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ABSTRACT

Deep core borings and deep open-cut excavations at numerous localities in the Tennessee Valley disclose the presence of solution cavities at considerable depths below the beds of the master-streams. At localities where structural and lithologic conditions are favorable, cavities are present at depths in excess of 100 feet below the water table. As the streams are flowing on bedrock in the deepest channels excavated at all localities, except four on the lower portion of the Tennessee River, it is concluded that the solution of carbonate rocks within the zone of saturation is a normal geologic process.

INTRODUCTION

Although much has been written within the past decade on the origin of limestone caverns, very little concrete information about the actual existence of solution cavities below the water table has been presented in the various papers. In 1937 the writer briefly called attention to the existence of cavities 100 feet below the bottom of the Tennessee River.[1] This paper presents the results of a more extensive study of subriver solution cavities in the Tennessee Valley.

SOURCES OF DATA

More deep drilling has perhaps been done on the Tennessee River system than on any other river system of comparable size in the world. Prior to 1933 the U.S. Army engineers did core drilling at numerous tentatively selected dam sites on the Tennessee River and its tributaries. Subsequent to 1933 intensive drilling has been done by the Tennessee Valley Authority at several additional localities in the area. A considerable amount of drilling has also been done along the rivers of the area by private organizations interested in dam sites and mineral deposits. As a majority of the localities in which drilling operations have been carried out are in calcareous rocks, an excellent opportunity is afforded for the study of the oc-

[1] Berlen C. Moneymaker, "Deep Solution Cavities in the Tennessee Valley Area," *Geol. Soc. Amer. Proc. 1937* (1938), p. 101.

currence, relative abundance, and size of solution cavities below the surface of ground water. The writer has studied several hundred thousand feet of core borings and entered numerous cavities penetrated by large-diameter drill holes and exposed in excavations. He has, in addition, studied a great number of drill records and actual cores made available by organizations other than the Tennessee Valley Authority.

The drill holes have ranged in diameter from an inch to 72 inches and in depth from a few feet to 646 feet. Although most of the exploration holes have been 2 inches in diameter, a great number of 6-inch holes have been drilled. The large-diameter holes, drilled for both exploration and foundation treatment, ranged in diameter from 30 to 72 inches, 36-inch holes predominating.[2] Most of the holes were carried to depths of less than 100 feet below the river bottom, but the deepest holes extend to 600 feet below river bed.

LIMITATION OF METHOD OF STUDY

The study of subriver cavities exposed in open-cut excavations and large-diameter drill holes is attended with as little difficulty as the examination of naturally exposed cavities. Inaccessible cavities known only from small-diameter drill holes are much less amenable to study. A small drill hole affords only a vertical dimension, which may or may not represent the maximum height of the cavity. Some of the apparently large cavities, especially those in calcareous sandstone and cherty limestone, are enlarged vertical joints and, although high, are narrow. Many of the apparently small cavities in essentially horizontal limestone are developed along bedding planes and, although low, have large lateral extension.

At all localities there are many more shallow holes than deep ones. Consequently, the deeper zones of cavitation are reached by comparatively few borings, and the average depths below river bed given in Table 1 are therefore weighted in favor of the shallower depths.

GENERAL GEOLOGY OF THE AREA

The Appalachian Mountain region is characterized by structurally complex crystalline rocks of pre-Cambrian and early Cambrian age.

[2] Berlen C. Moneymaker and Portland P. Fox, "Large-Diameter Core Drill for Geologic Exploration," *A.I.M.E. Tech. Pub. 1000* (1938).

These rocks are predominantly siliceous and consist mainly of gneisses, schists, quartzites, and mylonitized granites, although marble occurs in several localities.

TABLE 1

Locality Number	Rock Formation	Cavities			
		Maximum Vertical Thickness	Average Vertical Thickness	Maximum Depth below River Bed	Average Depth below River Bed
1	Fort Payne	64.1	1.6	222.7	80.6
2	Warsaw	1.6	0.9	79.2	9.4
3	Fort Payne	0.8	0.8	82.4	81.6
4	Fort Payne	3.5	0.8	45.9	26.0
5	Olive Hill	9.0	2.2	38.3	10.2
6	Fort Payne	8.3	2.2	32.0	12.7
7	Bangor (unrestricted)	7.3	1.0	46.3	13.8
8	Bangor (unrestricted)	32.4	3.0	110.8	36.5
9	Bangor (unrestricted)	17.0	2.1	58.2	17.6
10	Knox dolomite (undifferentiated)	4.0	1.0	32.0	20.5
11	Bangor (unrestricted)	18.0	1.7	102.5	31.9
12	Bangor (unrestricted), Fort Payne	4.5	2.0	40.7	13.8
13	Longview; post Longview	24.4	4.4	53.7	31.2
14	Chickamauga	53.8	2.8	110.0	30.1
15	Knox dolomite (undifferentiated)	31.6	4.5	165.1	52.2
16	Knox dolomite (undifferentiated)	6.7	2.1	37.6	13.5
17	Chickamauga	19.1	2.9	66.3	22.7
18	Chickamauga	2.5	1.4	16.4	8.3
19	Longview	1.7	0.9	27.1	19.0
20	Holston, Tellico	19.6	2.2	71.3	21.0
21	Tellico, Sevier	13.0	1.4	97.2	15.9
22	Copper Ridge	28.1	3.8	93.9	25.7
23	Lenoir	4.3	1.0	39.3	14.5
24	Copper Ridge	4.2	4.2	22.2	18.0
25	Copper Ridge, Chepultepec	6.6	3.6	25.6	18.6
26	Copper Ridge, Chepultepec	14.0	.....	28.0	.....
27	Copper Ridge	3.0	0.3	86.2	27.1
28	Murphy marble	10.0	5.0	75.0	20.0
29	Murphy marble	60.0	11.3	141.0	36.0
30	Murphy marble	21.0	3.5	75.0	25.6

The Appalachian Valley is a region of unaltered but highly deformed sedimentary rocks, mainly of early Paleozoic age. Calcareous rocks—limestone, dolomite, and shales—predominate, but sandstones and arenaceous shales are also present.

The Cumberland Plateau region is characterized by unaltered and

but slightly deformed sandstone and shale of the Pennsylvanian system. Mississippian limestones and chert immediately underlie the sandstone and shale and are exposed in the gorge of the Tennessee River west of Chattanooga.

The Central Basin and Highland Rim are characterized by slightly deformed calcareous rocks of middle and lower Paleozoic

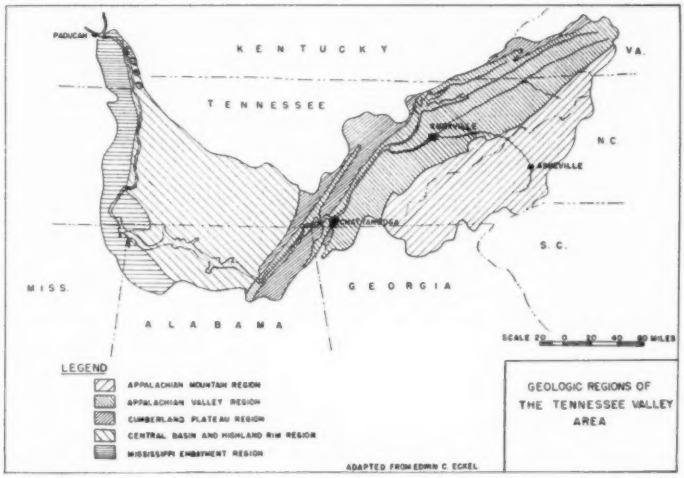

FIG. 1

systems. Thick-bedded and very pure limestones cover extensive areas, but cherty limestone and shale also have wide distribution.

The Mississippi embayment is an area of unconsolidated siliceous and argillaceous sediments, mainly gravel, sands, and clays of upper Cretaceous and Tertiary age. These deposits overlie the Mississippian limestone and other calcareous rocks of the middle and lower Paleozoic systems.

In each of these regions core drilling has been done in the investigation of dam sites or mineral deposits, or both, and in each of them carbonate rocks have been penetrated by drill holes. Where drilling into carbonate rocks has been done, cavities were invariably encountered at various depths below the bed of the river. The 30 localities at which deep drilling reveals subriver solution cavities are

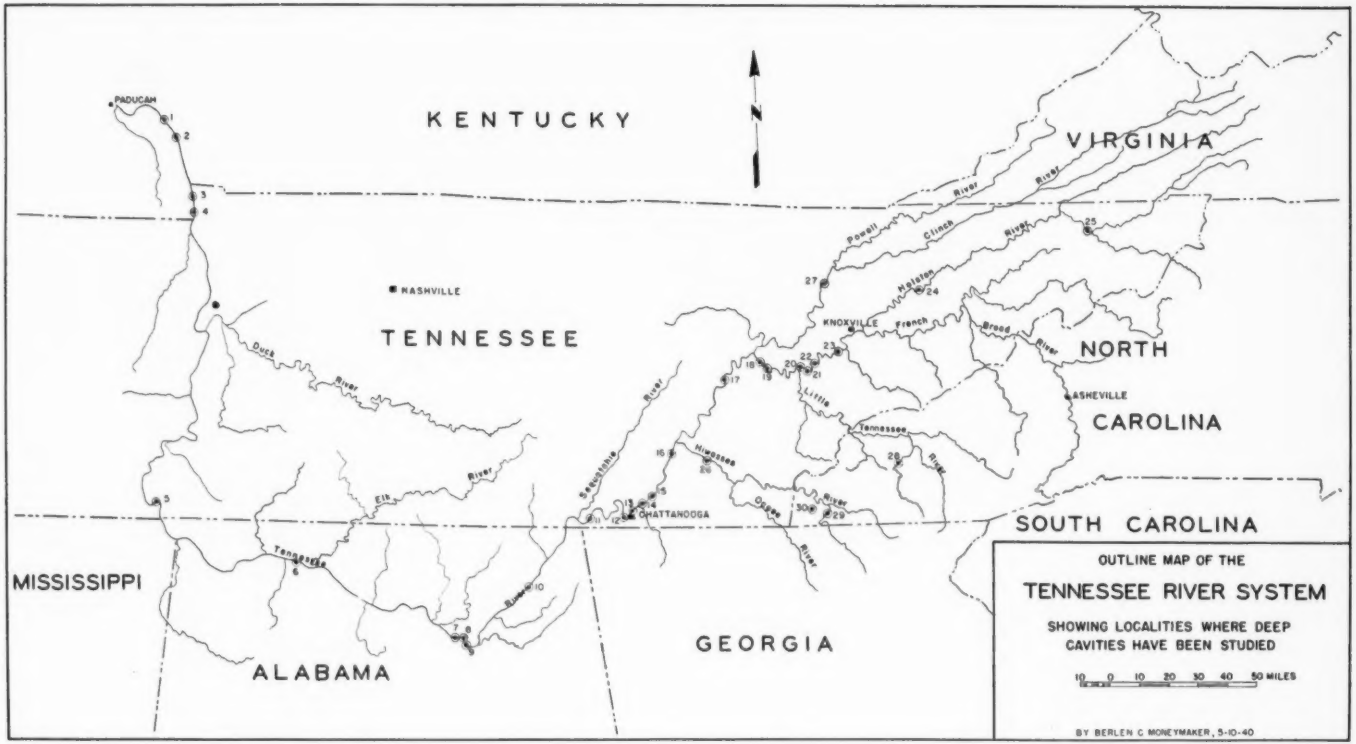
KENTUCKY
VIRGINIA
TENNESSEE
NORTH CAROLINA
SOUTH CAROLINA
GEORGIA
ALABAMA
MISSISSIPPI
PADUCAH
NASHVILLE
KNOXVILLE
CHATTANOOGA
ASHEVILLE
Duck River
Elk River
Tennessee River
Powell River
Clinch River
Holston River
French Broad River
Little Tennessee River
Hiwassee River
Ocoee River
OUTLINE MAP OF THE
TENNESSEE RIVER SYSTEM
SHOWING LOCALITIES WHERE DEEP CAVITIES HAVE BEEN STUDIED
10 0 10 20 30 40 50 MILES
BY BERLEN C MONEYMAKER, 5-10-40

FIG. 2

shown in Figure 2. Except Localities 1–4, inclusive, all these are in areas where the streams are now flowing on bedrock in the deepest channels excavated. It is thus clear that the cavities were formed by solution below river bottom and below the water table.

SUBRIVER CAVITIES IN THE VARIOUS GEOLOGIC REGIONS

APPALACHIAN MOUNTAIN REGION

Subriver solution cavities have been encountered at three localities (Nos. 28, 29, 30, Fig. 2) in the mountain region. These are in the steeply dipping and much fractured Murphy marble on the Nantahala, Hiwassee, and Nottely rivers, respectively. Both drilling and mining operations reveal that the marble is quite cavernous to considerably below the river bed. A total of 90 subriver solution channels, ranging in vertical dimension from a few inches to 60 feet and in depth below the elevation of river bed from a few feet to 141 feet, are known in the area. The average vertical dimension is about 11 feet, and about two-thirds of the cavities are within 35 feet of the elevation of the river bed.

APPALACHIAN VALLEY REGION

Solution cavities have been encountered below the river bed in at least fifteen localities (Nos. 12–26) in the Appalachian Valley portion of the Tennessee River basin. The geologic formations penetrated in these localities are shown in Table 1.

At each of these localities most of the holes drilled to elevations below that of the river bed encountered cavities, whether drilled from the river channel, flood plain, or higher ground. The maximum size and depth of the cavities below the river bed varied widely from place to place, depending largely upon lithologic and structural conditions. At some localities only a few holes were drilled as much as 50 feet below the river, consequently the maximum depth to which cavities extend was not determined with certainty. Other localities, however, were very thoroughly drilled, and the findings are sufficiently significant to merit special consideration.

Locality 14.—At Locality 14, the site of Chickamauga Dam, the Tennessee River crosses the great variety of thin- to thick-bedded limestones with thin interstratified layers of bentonite and shale,

known collectively as the Chickamauga limestone. The formation in this area is about 2,000 feet thick and includes all Ordovician strata above the Knox dolomite. It is subdivided by Fox[3] into the following units: Stones River (undifferentiated), Black River (Carter and Tyrone), Trenton (Hermitage and Cannon), and Richmond (undifferentiated). All the limestones of the Chickamauga are relatively pure and quite susceptible to solution. A major thrust fault with a stratigraphic displacement of about 5,000 feet brings the Chickamauga in contact with the Knox dolomite immediately upstream from Locality 14. As the Knox dolomite is thrust over the Chickamauga limestone, both formations are deformed for a considerable distance on either side of the fault. Many sharp folds and an even greater number of faults occur in the rocks of Locality 14. This structural deformation has facilitated the extensive development of cavities in the area.

Thousands of holes ranging from an inch to 72 inches in diameter have been drilled, and extensive excavations have been made at the locality. Both drilling and excavating have disclosed extensively developed cavities below the elevation of the river bed (Fig. 3). The lower limit of cavernous rock was, with few exceptions, controlled by the position of a thin bed of bentonite and shale near the base of the Cannon limestone. The position of this bed, which effectively protected the underlying rock from solution, varies considerably from place to place as a result of folding and faulting. The deepest cavity in the area extends downward to 110 feet below the elevation of the river bed. Although the maximum vertical dimension of a subriver cavity here was 54 feet, the average of approximately 2,000 such cavities is less than 3 feet.

It was not possible to determine the channel pattern with accuracy, but excavations and ground-water studies indicated that some of the larger channels are hundreds, if not thousands, of feet long, of very irregular trend, and of variable dimensions. Where intersected by large holes or excavations, open or partly open channels discharged copious quantities of water.

Locality 15.—This is in the extensively fractured Knox dolomite a short distance upstream from the Knox Chickamauga contact fault.

[3] Portland P. Fox, "Foundation Exploration and Geologic Studies at Chickamauga Dam," *Proc. Amer. Soc. Civil Engineers*, Vol. LXVI, No. 3 (1940), pp. 465–68.

The lack of impervious and insoluble beds in the Knox dolomite enabled solution to progress to greater depths than in the near-by Chickamauga limestone. The comparatively few holes at this locality reveal the existence of cavities to a depth of 165 feet below the river

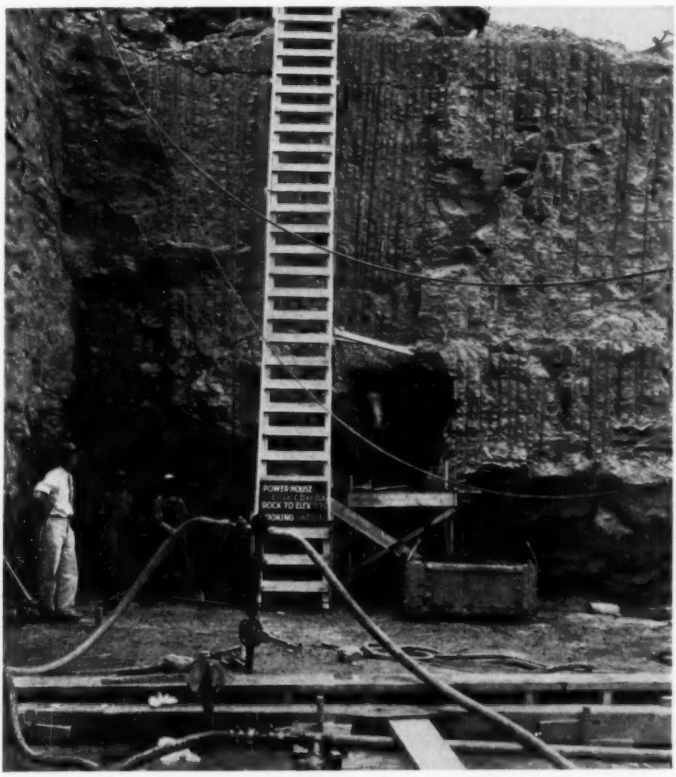

FIG. 3.—An open cavity 35 feet below the bed of Tennessee River at Locality 14

bed. The largest individual cavity has a vertical dimension of 31.6 feet and extends to 79 feet below the river. The average thickness (vertical dimension) of the cavities at this locality is 4.5 feet, and the average depth below the river bed is 52 feet.

Localities 20 and 21.—Here bedrock consists of the Holston "marble" (coarsely crystalline limestone), Tellico sandstone (cal-

careous sandstone and sandy limestone), and Sevier shale (calcareous shale and limestone). Although the limestone beds in the Sevier are invariably cavernous, the larger and deeper cavities are in the Holston and Tellico. In both of these formations the cavities are deep, narrow channels representing the enlargement by solution of steeply inclined joints. Cavities extend to 71 and 97 feet below the elevation of river bed at Localities 20 and 21, respectively.

Locality 22.—Geologically, Locality 22 is quite similar to Locality 15. It is in the Knox dolomite in close proximity to a major fault, and the rock is highly fractured. A relatively small amount of drilling disclosed the presence of cavities to 94 feet below the river bed. The largest cavity encountered, 28 feet in vertical dimension, was 35 feet below the river bed.

Locality 27.—This, the site of Norris Dam, is in the Knox dolomite in essentially the same strata as Localities 15 and 22. The rocks are almost horizontal and are entirely unfaulted. Both drilling and excavation disclosed numerous small subriver cavities. These openings represent the enlargement of bedding planes and joints. The average thickness of over 300 openings attributable to solution is 0.3 feet, the largest one being 3 feet thick. The deepest cavity here was 86 feet below the bed of Clinch River.

CUMBERLAND PLATEAU REGION

All the localities within the confines of the Cumberland Plateau physiographic province (Localities 7–12, inclusive, Fig. 2) are in sub-Carboniferous limestones exposed where the Tennessee River has cut through the coal measures. Lithologically, this region resembles the Appalachian Valley region and may be treated as its continuation (Fig. 1). With one exception, Locality 10, the localities are in the pure limestones of the Bangor (unrestricted) limestone.

Locality 11.—This, the site of Hales Bar Dam, is in Mississippian limestone, which may be subdivided into the following formations: Ste. Genevieve limestone, Gasper limestone, Golconda shale, Hartselle sandstone, and Bangor (restricted) limestone. Numerous drill holes and open excavations have revealed that solution channels are extensively developed in the rock underlying the river.[4] The largest

[4] J. A. Switzer, "The Power Development at Hales Bar," *Tenn. Geol. Surv.* ("Resources of Tennessee"), Vol. II, No. 3 (March, 1912), pp. 92–93.

channel known to exist beneath the river has a vertical dimension of 18 feet, and the deepest extends to a depth of 102 feet below the channel of the river.

Locality 7.—This, the site of Guntersville Dam, is also in the Bangor (unrestricted) limestone. The geologic structure is quite simple, as the strata are essentially horizontal and unfaulted. The rock is cut by three sets of joints which have played an important role in subriver solution. The stratum forming the floor of the river

FIG. 4.—An extensively dissolved stratum forming the floor of the river channel at Locality 7.

over a wide area has been reduced by solution along joints to numerous isolated blocks (Fig. 4). As at Chickamauga Dam, the depth to which solution has progressed is controlled by a shale bed which has effectively protected the underlying rock.[5] The maximum depth of solution is 46 feet below the river, and the largest cavity encountered had a vertical dimension of 7.3 feet. At Locality 8, a short distance upstream, the shale does not occur immediately below the river, and solution cavities extend to a depth of 111 feet below the river floor.

[5] Robert M. Ross, "Foundation Exploration and Geologic Studies at Guntersville Dam," *Proc. Amer. Soc. Civil Engineers*, Vol. LXVI, No. 3 (March, 1940), pp. 424–27.

CENTRAL BASIN AND HIGHLAND RIM REGION

The only locality in this region is No. 6, the site of Wheeler Dam. Bedrock immediately below the river is the Fort Payne chert. Because of its highly siliceous character and the scarcity of joints, this rock was not very cavernous. The deepest cavity encountered was 32 feet below the bed of the river, and the largest one was 8.3 feet from roof to floor.

MISSISSIPPI EMBAYMENT REGION

In the Mississippi embayment region, subriver cavities have been encountered in drill holes at five localities (1–5, inclusive, Fig. 2). Bedrock is the Fort Payne chert at localities 1, 3, and 4, Warsaw limestone at Locality 2, and Olive Hill at Locality 5. Bedrock is nearly everywhere concealed by upper Cretaceous (Ripley, Tuscaloosa, and Eutaw) and Tertiary (terrace gravel, loess, and river alluvium), except along the Tennessee River, where the underlying Mississippian limestone has been uncovered in many places.

In the lower portion of its course the Tennessee River is flowing on thick fluviatile sediments; in Pleistocene time the river carved a deep valley which has subsequently been partially filled with alluvium.[6] At Locality 1, the buried channel is at an elevation approximately 70 feet lower than that of the present river channel. In view of this fact it is apparent that only the cavities occurring in this area at elevations lower than that of the old channel throw any light on solution below the level of ground water. According to Rhoades,[7] "the ground water table stands at the present time at 302 feet above sea level near the river and rises to about 315 feet near the outer edges of the flood plains. Formerly, however, when the rivers occupied the deeper parts of their buried valleys, the water table was doubtless depressed to about elevation 230 feet. Deep solution can only be discussed with reference to this former depressed water table; otherwise, solution effects which seem to be deep as compared to the present river may have been formed at normal elevations relative to the historic, lower, river surface."

Drilling at Locality 1, the site of Kentucky Dam, has disclosed the existence of hundreds of cavities below elevation 230 feet. The

[6] Roger F. Rhoades, "Profiles of the Buried Valleys of the Cumberland, Ohio, and Tennessee Rivers" (abstr.), *Geol. Soc. Amer. Proc. 1936* (1937), p. 97.

[7] Personal communication, December 21, 1937.

deepest cavity encountered to date is at elevation 68 feet—162 feet below the former depressed water table and 234 feet below the present water table. Although the largest known cavity has a maximum vertical dimension of 64.1 feet, the average vertical thickness of about 1,300 cavities below the elevation of the present river bed is only 1.6 feet.

FILLED CAVITIES

A large percentage of the subriver solution cavities in the Tennessee Valley are partially or completely filled. All filled and partially filled cavities contain residual clay, silt, sand, and gravel, or mixtures of these materials. Many of them also contain water-worn cobbles, and a few of them contain rounded boulders up to a foot in diameter. In general the smaller and deeper cavities are filled with clay or silt, and the larger cavities and those near the top of bedrock contain coarser materials. A large cavity below the flood plain of Nottely River at Locality 30 contained water-worn boulders of quartzite up to a foot in diameter. At Locality 14, a cavity under the flood plain 300 feet away from the river and 50 feet below the top of bedrock contained cobbles and a few boulders up to a foot in diameter.

There is no apparent relation between either the depth or the size of cavities and the presence or absence of filling. Large cavities may be open or filled regardless of depth below the elevation of river bed, and this is equally true of small cavities. In a deep drill hole, which penetrates numerous cavities ranging in vertical thickness from a few inches to many feet, open and filled cavities commonly occur in irregularly alternating series, irrespective of either depth or size.

SIGNIFICANCE OF SOLUTION WITHIN THE ZONE OF SATURATION

In many of the localities listed in Table 1 drill holes and excavations have afforded a means for the study of cavities above the elevation of the river bed both below and above the water table. A comparison discloses that, within the zone of saturation, cavities are invariably more numerous and of greater average size above the elevation of river bed than below it. Cavities above the water table are found, in turn, to be more numerous and of greater average size than those within the zone of saturation. These observations suggest that, although the development of caverns may be begun within the zone of saturation, the major portion of enlargement takes place

above the water table. This conclusion, however, is not yet capable of demonstrable proof. The water table has occupied successively lower positions in response to the downcutting of the streams; consequently each level above the present water table has occupied a position which, in turn, was below, coincident with, and above the water table. All solution cavities above the present water table may have been associated with previous water tables. It is not possible to determine precisely how much of the enlargement is attributable to the present rather than to the past water-table relationships.

Solution of rock below the bed of a large river undoubtedly plays an important role in the downcutting of the stream's channel. From Knoxville, Tennessee, to Guntersville, Alabama, the channel of the Tennessee River consists of an alternating series of deep pools and shoals. A study of that section of the river reveals that the pools are, without exception, in soluble rocks and the shoals are in insoluble rocks.[8] As the rocks forming the shoals are in many cases less resistant to abrasion than those forming the pools, it is believed that solution is largely responsible for the pools. This belief is supported by the cavernous character of the rocks underlying the pools, as shown in core borings at several localities.

CONCLUSION

A study of deep drill holes and deep open cuts at numerous widely separated localities in an extensive area of variable geologic character reveals that the development of solution channels in limestone, dolomite, marble, and other carbonate rocks within the zone of saturation is a normal geologic process. Channels are most numerous, largest, and extend to greatest depths in zones of deformation and especially where the movement of ground water is unimpeded by insoluble or impervious beds.

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[8] Moneymaker and Fox, "Origin of Deep Pools in the Channel of Tennessee River," *Bull. Geol. Soc. Amer.*, Vol. L, No. 12, Part II (1939), pp. 1923–24.

STUDIES FOR STUDENTS

RECOGNITION OF FAULTS IN SOUTHERN CALIFORNIA

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ABSTRACT

In this paper evidence is presented to show that faults are more numerous and geomorphically more important than various geologists believe. Fifteen criteria for the recognition of the faults, especially those in crystalline rocks, are discussed. Many examples are given.

GENERAL STATEMENT

When it comes to a consideration of the number of faults in southern California and the evidence for their existence, geologists are more or less divided into two groups. In one group are those who believe that a fault should not be mapped unless the actual fault can be seen nearly everywhere or unless there is positive stratigraphic evidence for it. Many of these workers are not very familiar with fault problems in crystalline rocks. In another group are those who recognize much more faulting on the basis of still other criteria, particularly in areas of crystalline rocks. The writer belongs in the second group, and he will discuss various criteria of faulting in addition to actual dislocation of definitely determined amount based upon stratigraphic evidence.

In a lengthy discussion R. D. Reed[1] has urged caution and the need of careful field work in locating faults, stating that too many faults have been assumed to exist, often on the basis of not more than physiographic evidence or general field reconnaissance. In support of his view he cites a number of cases where supposed faults failed to be proved with more detailed field work. He is definitely conservative about the number of faults in southern California. The writer agrees with Reed that caution should be exercised and that a

[1] *Geology of California* (Tulsa: Amer. Assoc. Pet. Geol., 1933), pp. 48–59.

fault should not be mapped unless there is substantial field evidence for it. Detailed regional studies are, however, constantly bringing to light many heretofore unknown faults in southern California. Thus, several faults with displacements varying to some hundreds of feet have been proved very recently in both the Huntington Beach and Wilmington oil fields in the Los Angeles Basin. Such, in general, has been the writer's experience where he has done detailed field work in many parts of southern California. In this connection the writer may mention a personal communication from Reed stating that there is no fault through the depression between Coyote and Fish Creek Mountains west-northwest of El Centro as shown on a map by the writer. He is partly correct about this because, owing to an error in drafting, this fault is shown somewhat too far north. But what he does not say is that, instead of a single fault through this depression, there are two definitely proved faults—one on each side of the depression. Furthermore, on R. D. Reed and J. S. Hollister's recent tectonic map[2] of a part of southern California various fairly important, well-proved faults are not shown, as, for example, along the southwestern San Gabriel and Verdugo Mountains north of Los Angeles.

As another example of conservatism in recognizing faults and their geomorphic influence, mention may be made of certain statements by Bailey Willis on the Peninsular Mountains between the seacoast and the Imperial Valley. Thus he says[3] that "in a recent reconnaissance" from seacoast to Salton Basin he "failed to find evidence of extensive faulting," that "erosion and warping were chiefly responsible for the scenic character of the mountains," and that there is need for "much more thorough investigation, including mapping of formations." He disagrees with the present writer, who has emphasized the importance of faulting. But one may wonder whether such statements, without any discussion of the evidence presented elaborately by the writer,[4] are a fair disposal of the matter. The writer's argu-

[2] *Structural Evolution of Southern California* (Tulsa: Amer. Assoc. Pet. Geol., 1936, map in pocket).

[3] "San Andreas Rift in Southwestern California," *Jour. Geol.*, Vol. XLVI (1938), pp. 1047 and 1049–50.

[4] "Geomorphology of the Southern Peninsular Range of California," *Bull. Geol. Soc. Amer.*, Vol. XLVI (1935), pp. 1535–62.

ments should be discussed categorically. In the second place, a quick reconnaissance across this extensive region is by no means a sufficient basis for the conclusions reached by Willis. The present writer spent many weeks in the region studying the geomorphology, giving particular attention to the influence of faulting upon the origin of the land forms. He also published a detailed geologic map[5] of a wide strip across the mountains.

CRITERIA FOR THE LOCATION OF THE FAULTS

Many criteria for faulting can be applied in southern California. Before discussing these, it should be definitely understood that they are of varying value. A single criterion may be conclusive evidence of faulting, another may be of little value by itself. Where several otherwise individually inconclusive criteria can be applied and proper precautions have been taken as to the nature and structure of the rocks (other than faulting), a case for faulting is greatly strengthened.

I. FAULT SURFACES ACTUALLY VISIBLE

Some of the many places where important fault surfaces have actually been seen by the writer are as follows: San Gabriel fault east of Newhall; a fault in the San Gabriel zone in Little Tujunga Canyon; Sierra Madre fault in Little Tujunga Canyon; Soledad fault near Russ Siding in Soledad Canyon; Burbank fault near Burbank; Fish Creek fault on south side of Fish Creek Mountains; Coyote fault on north side of Coyote Mountains; fault on south side of Calico Mountains at Calico; and fault at west base of Dead Mountains northwest of Needles. It is interesting to note that, in all cases cited, sedimentary rocks are faulted against crystalline rocks where conditions were especially favorable for sharpness of fracturing. In southern California, however, as in most other regions, sharply defined surfaces of important faults are seldom actually visible for various reasons. Also, clean-cut fault surfaces often do not occur at all, among other reasons because of the prevalence of horizontal shearing and crushing of the rocks, especially the crystalline rocks.

[5] "A Geologic Section across the Southern Peninsular Range of California," *Calif. Jour. Mines and Geol., Rept. 31 of State Mineralogist* (1935), pp. 115–42.

II. VERTICALLY DISLOCATED STRATA

This is a very important and often definite criterion for location of faults which transect strata, particularly in the large region from the Los Angeles Basin northwestward. In addition to the usual surface stratigraphic evidence, subsurface evidence from oil-well records has often proved the existence of faults as, for example, in the Wilmington and Huntington Beach oil fields, where buried faults have been proved. Further discussion of this well-known criterion for locating faults is not necessary here.

III. VERTICALLY DISLOCATED OLD EROSION SURFACES

It is generally agreed that, in late Tertiary or early Quaternary time, all or nearly all the area of southern California was in a topographic condition varying from early old age to a peneplain. Large and small remnants of that surface, little affected by subsequent erosion and often covered with residual soil, are well preserved in many places and at various altitudes. In many other places, however, the old surface has either been destroyed by erosion or it lies buried under Quaternary alluvium. The writer believes that relatively long, straight scarps, separating such old erosion surfaces sharply at different levels, are almost as good as positive stratigraphic evidence of faulting. Excellent examples are to be found in the southern Peninsular Range of California. Thus, the Laguna Mountains old surface at an elevation of about 6,000 feet is sharply separated by a long, steep scarp from the McCain's Plateau old surface 2,000 feet lower. The former is distinctly tilted, while the latter larger surface is essentially level. Each is largely covered with deep residual soil, and each has scattering low monadnocks. C. O. Sauer's argument[6] that similar old-age surfaces just northwest of these are "Primarrumpfe" produced by erosion almost entirely while they were being uplifted rather than before they were uplifted is answered by the writer[7] in his study of the geomorphology of that region.

[6] "Landforms in the Peninsular Range of California as Developed about Warner's Hot Springs and Mesa Grande," *Univ. Calif. Pub., Bull. Dept. Geog.*, Vol. III (1929), p. 248.

[7] "Geomorphology of the Southern Peninsular Range of California," *op. cit.*, pp. 1550–52.

Some other examples of remnants of the old erosion surface now high up on mountains are as follows: Big Bear Valley region (about 7,000 feet) in the San Bernardino Mountains; Barley Flats and Pine Flats (about 6,000 feet) in the San Gabriel Mountains; Pinyon (about 4,000 feet) in the northern part of the Santa Rosa Mountains; and several wide alluvium or "wash"-covered erosion surfaces near the summit of the Ord Mountains south-southeast of Barstow. The western slope of the Santa Ana Mountains fault block represents a tilted and partly dissected portion of the once general old surface. The Verdugo Mountains block, north of Los Angeles, is sharply divided into three sections separated by scarps hundreds of feet high, the summit of each section marking part of the old erosion surface. The extensive area called the "Perris peneplain" (altitude about 1,500 feet) represents part of the once widespread, old-age surface relatively little uplifted and lying between the much more greatly uplifted San Jacinto and Santa Ana block mountains adjacent to it on two sides.

IV. VERTICALLY DISLOCATED VOLCANIC ROCKS

Where Cenozoic volcanic rocks lie in sharp contact, without matching, along straight lines for a mile or more, there is strong evidence that the contact is a fault. Where Cenozoic volcanic rocks form sharp, straight contacts for a mile or more against plutonic, metamorphic, or sedimentary rocks, faulting is strongly suggested, particularly where the volcanics lie at the same level or lower than the adjacent older rocks. Fine examples involving these criteria are: on the northwestern and southern sides of the San Gabriel Mountains in the general vicinities of Acton and Glendora, respectively; in the Newberry Mountains east-southeast of Barstow; in the Whipple Mountains near Parker Dam; and in the western Santa Monica Mountains.

Otay, Jamul, and San Miguel Mountains stand out boldly above the surrounding country in the Peninsular Range east-southeast of San Diego. Because the rocks of these mountains are largely volcanic, it might be argued that they have resisted erosion better than the neighboring rocks, thus allowing these masses to stand out boldly. This cannot, however, be more than part of the explanation, be-

cause other large areas of the volcanics, such as west and south of Mount Helix, lie at far lower levels. The distinctly triangular, quadrangular, and pentagonal shapes of the mountain masses and associated valleys also indicate strongly that they are fault blocks rather than results of stream erosion unaffected by structural control and uplift.

Dislocated, lava-capped old erosion surfaces are easily recognized in many parts of the Basin and Range province, but good examples do not seem to be common in southern California. A fine case in point is where lava-capped, reddish, late Cenozoic gravels, separated vertically by 2,000 feet, lie at the top and the bottom of the scarp at the eastern base of the Peninsular Mountains a few miles from the Mexican border.

V. HORIZONTAL DISLOCATIONS (OFFSETS)

Criteria for horizontal displacements, ranging probably to many miles, are remarkably evident in southern California, particularly along the San Andreas and Garlock faults. In view of the fact that this important matter has already been rather fully discussed by the writer,[8] further comments are unnecessary here.

VI. FAULT CRUSH-ZONES

Fault breccia and crush-zones afford positive evidence of faulting. Such phenomena are particularly common and significant within the bodies of relatively hard and brittle crystalline rocks, especially at bottoms of various relatively straight canyons or in saddles or cols, especially in cases where the latter show alignment. Fault crush-zones are commonly from a few feet to 100 feet or more in width, their size in a general way indicating the magnitude of the faulting. Crush-zones are most conspicuously developed in crystalline rocks where much horizontal movement has occurred along faults under conditions of great pressure and shearing stress. Such conditions were fulfilled on a grand scale along the San Andreas fault as plainly exposed in a road cut where this fault crosses Cajon Canyon northwest of San Bernardino. A zone of pre-Cambrian foliated rocks, several hundred feet wide, is there so highly crushed that much of the

[8] "Some Features of Faulting in Southern California," *Jour. Geol.*, Vol. XLVIII (1940), pp. 397–401.

rock is pulplike. Along the same fault in the northeastern San Gabriel Mountains, a little west of Big Pines Camp, granitic rocks have been ground to a white "rock pulp" in a zone about 100 feet wide. Hundreds of other crush-zones have been observed by the writer in southern California, and a few good examples may be mentioned as follows: in the San Gabriel Mountains, in San Gabriel Canyon 1½ miles from its mouth, several places along West Fork San Gabriel River, and in Big Tujunga Canyon west of Hoyt Ranch; various places along the western side of the Little San Bernardino Mountains; and various places along faults in the Peninsular Range of southern California.

VII. ALIGNMENT OF SADDLES

Saddles produced by erosion along faults or fault crush-zones occur in various places. Detailed field work by the writer has brought to light some excellent examples in the western San Gabriel Mountains. Alignment of saddles may be produced by erosion irrespective of fault control, but in such cases belts of otherwise weak rock are the usual control. If a line of saddles cuts across rocks without regard to their kinds or structures, fault control is a fair presumption. Lack of exposures in many cases makes definite decisions difficult or impossible. An example to the contrary was observed by the writer where a wide crush-zone, completely concealed in a saddle, was exposed in the Colorado River Aqueduct Tunnel under that saddle in the eastern side of the Iron Mountains.

VIII. ZONES OF EXCESSIVE JOINTING WITH SLICKENSIDES

In view of the fact that the significance of this criterion as an evidence of faulting is similar to that of fault crush-zones just discussed, brief consideration only is necessary. Zones of excessive jointing near known faults, particularly in crystalline rocks, are common occurrences. Long, relatively straight zones of this kind often lead to the finding of important faults or fault zones. The writer has observed all types of structure from long, relatively straight and narrow zones of more than usual jointing with little or no slickensiding on the joint faces, through zones of excessive jointing with slippage on joint faces very common, to real faults, either clean cut or with crush zones.

Small slippages, in the same general direction, on numerous closely spaced joint faces may result in a distributive fault of considerable magnitude or at least contribute to the magnitude of a fault zone. Only two examples will be mentioned. Thus, very conspicuous zones of excessive jointing, accompanied by well-defined fault crush-zones, occur in diorite along the face of the Elsinore fault scarp near the international border. In the other case there is a wide zone of closely spaced, steep-dipping joints in monzonite on a ridge in the Eagle Mountains. The structure is accentuated by weathering and erosion. Hundreds of feet underneath in the aqueduct tunnel the writer saw the same structure but with slickensiding plainly visible on nearly every joint face of the fresh rock.

IX. STRAIGHT SCARPS CUTTING ACROSS VARIOUS ROCKS AND STRUCTURES

The writer believes that long, straight, conspicuous scarps cutting across country, irrespective of kinds and structures (other than faulting) of the rocks, are in many cases strong indicators of faulting. This criterion is greatly strengthened when taken in conjunction with the next one (No. X) because of the elimination of stream erosion as a possible cause of such scarps. Good illustrations are various northeasterly trending scarps which cut squarely across both igneous and metamorphic rocks of various kinds and the foliation of the metamorphics in the southern Peninsular Range of California. Most conspicuous of all is the long, high scarp forming the southeastern boundary of the Laguna Mountains. Some good examples occur in the western San Gabriel Mountains and in various other localities.

X. PROMINENT SCARPS CUTTING ACROSS IMPORTANT DRAINAGE COURSES

In many places important streams of a region cut squarely across long, prominent, relatively straight scarps and, where the nature and structure of the rocks can be eliminated as a cause of such scarps, the presumption favors their fault origin. The cross-cutting streams could not have produced such scarps. Two important north-northwesterly trending scarps, cross-cut by several important streams in the northeastern part of the Cuyamaca quadrangle in the Peninsular Range, are good illustrations of this principle. Numerous other examples occur in southern California. This criterion is more conclusive when applied in conjunction with No. IX, already mentioned.

XI. LONG, STRAIGHT STREAM COURSES CUTTING ACROSS VARIOUS ROCKS AND STRUCTURES

Streams of this kind are very common in southern California. Some good examples of such streams, the courses of which are definitely known to be controlled by fault zones of weakness, are as follows: Lone Pine Creek and West Fork San Gabriel River in the eastern and southern San Gabriel Mountains, respectively; Mill Creek in the southern San Bernardino Mountains; and several streams, particularly Cottonwood Creek, in the southern part of the Cuyamaca quadrangle in the Peninsular Range. In many cases where positive evidence of faulting is lacking but where long, straight stream courses are not due to superimposition and are uninfluenced by kinds of rocks or structures other than faulting, the presumption favors fault control.

XII. LONG, STRAIGHT SCARPS IN RELATION TO ADJACENT ALLUVIAL CONES AND PLAYA BASINS

As emphasized by W. M. Davis, very small or nonexistent alluvial cones along the base of a straight, high scarp in the desert are a good indication of recent faulting insufficient amount to carry the alluvial material down (by faulting) almost if not quite as fast as it accumulates at the base of the scarp. If the lowest part of the bolson, i.e., the *playa*, lies against the scarp, it is further evidence of active or recently active downfaulting at the base of the scarp. The best examples known to the writer in California are along the base of the long, steep scarp facing Deep Spring Valley and along the base of the scarp forming the eastern face of the southern part of Death Valley. Two cases may be mentioned. One is on the north side of Newberry Mountain east-southeast of Barstow, and the other is on the south side of the Eagle Mountains. In each case a high, steep scarp has a well-developed *playa* at its base but only slightly built-up alluvial cones. Positive evidence for faulting is lacking in the rocks, but the indirect evidence strongly favors the fault origin of the scarps.

XIII. FLAT-FLOORED VALLEYS BOUNDED BY HIGH, STRAIGHT SCARPS

A number of sharply defined, nearly flat-floored valleys, bounded by straight, steep scarps hundreds of feet high on one or more sides, exist irrespective of kinds and structures (other than faulting) of the

rocks in the Peninsular Range of southern California. An excellent case in point is San Felipe Valley in the Ramona quadrangle. C. O. Sauer[9] has argued that this valley was produced by denudation and erosion and that only one fault passes lengthwise through the middle of the valley. The writer[10] has answered this argument at length by contending that this valley and others similar to it in the Peninsular Range are essentially grabens bounded by a fault on each side. In view of the fact that this matter has been published at length, it seems unnecessary to repeat it here.

XIV. DISTURBED STRATA ADJACENT TO SCARPS

Definitely known cases of deformation of late Tertiary or early Quaternary strata adjacent to fault blocks are fairly common. A number of good examples have already been mentioned. Where such strata are strongly deformed in comparatively narrow belts along bases of straight scarps, particularly those of crystalline rocks, and faults cannot be actually seen or positively proved, the evidence strongly favors faulting.

XV. ALIGNMENT OF SPRINGS

Springs often occur along faults in southern California. In many cases lines of springs, often hot springs, occur along faults or fault zones. Good examples are Eden, Gilman, Soboba, and Ramona Hot Springs along the San Jacinto fault zone on the northwestern side of the San Jacinto Mountains and Glen Ivy and Murrietta Hot Springs along the Elsinore fault zone near Elsinore. Waters of several springs emerge from the fault zone on the north side of the San Bernardino Mountains. Other examples could be cited. When other evidence leaves considerable doubt as to the existence of a fault, an alignment of springs, particularly along the base of a long, straight scarp, strengthens the case for faulting.

XVI. ALIGNMENT OF SMALL SCARPS, RIDGES, AND DEPRESSIONS

It is well known that lines or narrow zones of small scarps, ridges, and depressions (including sag-ponds) commonly occur along faults, especially along important shear zones. Thus, as stated by L. F.

[9] *Op. cit.*

[10] "Geomorphology of the Southern Peninsular Range of California," *op. cit.*, p. 1552.

Noble,[11] the San Andreas fault throughout southern California "is marked by a curiously straight and almost continually traceable chain of scarps, ridges, and troughlike depressions, most of which involve Quaternary alluvial deposits and thus afford clear and unmistakable evidence of recent earth movements." Fine examples may be seen along the great fault within 25 or 30 miles on either side of Palmdale. Similar features mark parts of the Garlock fault, particularly west of Randsburg. Examples along other faults could be mentioned.

GEOMORPHIC INFLUENCE OF THE FAULTING

With few, if any, important exceptions the principal mountains of southern California are direct results of the Quaternary faulting which caused a general breaking-up of the late Tertiary–early Quaternary old-age surface. There are so many widely scattered, well-preserved remnants of that old surface, now at various altitudes because of subsequent faulting, particularly in the western part of southern California, that it was quite certainly very widespread. Proof that this old-age surface was broken up in late Cenozoic time lies in the fact that late Tertiary and Quaternary sediments have been in many places involved in the faulting.

The highest and most conspicuous mountains very generally lie adjacent to the greatest faults. A number of the principal ones, with maximum altitudes in feet, may be listed as follows: San Bernardino (11,485), Little San Bernardino (5,400), San Gabriel (10,080), Northern Ventura County (8,826), and Tecuya (7,178) Mountains adjacent to the San Andreas fault; the San Jacinto (10,805) and Santa Rosa (8,705) Mountains adjacent to the San Jacinto fault; Santa Ana (5,680), Aguanga (6,126), Laguna–Cuyamaca (6,515) Mountains adjacent to the Elsinore fault; and the Sierra Madre–San Rafael (6,589) Mountains adjacent to the Nacimiento fault. As shown in another paper by the writer,[12] great regional pressure from the south, accompanied by both vertical and northwesterly horizontal movements, caused the uplift of these mountains.

[11] "Southern California," *Guidebook 15, XVI Internat. Geol. Cong.* (1932), p. 11.

[12] Miller, "Some Features of Faulting in Southern California," *op. cit.*, p. 417.

It is significant that a number of the mountains just mentioned lie in a north-south belt from the San Bernardino to the Laguna Mountains and adjacent to the important bends in the San Andreas, San Jacinto, and Elsinore faults. No doubt extra-great pressure from the south in this belt produced both the bends in the faults and the accompanying high mountains. There is a similar relationship of especially high mountains to the great bend in the San Andreas fault near its junction with the Garlock fault. The lofty San Gabriel Mountains are exceptional in that they do not lie adjacent to a great bend in a prominent fault, but, as already shown, they owe their existence to the upsqueezing of a large wedge-shaped block, between the San Andreas and San Gabriel–Sierra Madre faults, by strong pressure from the south.

The south end of the Sierra Nevada Mountains (maximum altitude, 8,440 feet) and the Tehachapi Mountains (maximum altitude, 7,994 feet) are more related genetically to the Sierra Nevada and Garlock faults than to other important faults.

With few exceptions, mostly within 25 or 30 miles of the San Andreas fault, the extensive desert region lying between the San Andreas fault, Garlock fault, and the Colorado River is much less profoundly faulted than the rest of southern California, and the mountains are usually not so high. This is probably because the great regional pressure from the south met with such resistance that only relatively near the San Andreas fault were the horizontal (shearing) and vertical movements very effective. Even in this region, however, there was enough widespread fracturing and block faulting to produce all, or nearly all, the numerous important mountain masses of the region. In the eastern part of this extensive desert region, the Basin and Range province type of faulting seems to prevail.

Differential movements have caused many large and small relatively depressed regions such as the western part of the great Mohave Desert wedge, the large Salton Sea–Imperial Valley graben, the large graben between the San Jacinto and Santa Ana Mountains, and Ridge Basin graben between the San Andreas and San Gabriel faults at and near their junction. The last-named graben, as already shown, sank fully $4\frac{1}{2}$ miles between Middle Tertiary and Middle Pleistocene times.

Tilting of fault blocks has also been fairly common, excellent examples being the Santa Ana, San Gabriel, and Laguna Mountains, all downtilted to the southwest, and the Santa Rosa Mountains strongly downtilted to the northeast. It is interesting to note that the southern part of the Sierra Nevada,[13] unlike the main part of the range, shows little or no downtilt to the west.

In numerous cases fault scarps have developed across stream courses. In most cases the downthrow has been on the downstream side of the fault, thus speeding up erosive activity locally. In rarer cases the downthrow has been on the upstream side of the fault, thus interfering with, or locally slowing down, its erosive activity. If the stream maintains its course through such a rising scarp, it is, of course, an antecedent stream. A very fine illustration of the last-named principle is Morongo Valley and vicinity on the eastern side of the San Bernardino Mountains. This narrow valley, about 10 miles long, has plainly been produced mainly by faulting resulting in a scarp hundreds of feet high along the southeastern side of the valley. Each of three streams (Big, Little, and Dry Morongo creeks) flows directly across the floor of the valley and suddenly enters a deep, narrow gorge in the fault scarp and thence continues its course for miles. These are undoubtedly antecedent streams. The interested reader should consult the southeastern corner of the U.S.G.S. San Gorgonio quadrangle for a good map of this region.

Another case of what is very probably an antecedent stream is that of the Santa Ana River, which suddenly enters Santa Ana Canyon which has been cut through the Santa Ana Mountain block as the latter rose across the course of the river.

Mountain fronts, representing fault scarps modified by erosion, are very common throughout southern California. These are usually very steep, relatively straight, ranging in height to 10,000 feet and in length to scores of miles.

The question of the position of faults in relation to mountain fronts is important. It is the writer's belief that by far most of the long, straight, high, steep scarps in southern California are results of faulting, particularly those which are at least 5 or 10 miles long,

[13] W. J. Miller, "Geologic Sections across the Southern Sierra Nevada of California," *Univ. Calif. Pub., Bull. Dept. Geol. Sci.*, Vol. XX (1931), p. 334.

at least hundreds of feet high, and developed irrespective of rock types and structures (other than faulting). In such cases the faults lie at, or very close to, the bases of the scarps. Numerous definitely known examples occur throughout southern California. When fault activity along the base of such a scarp dies down greatly or ceases, the mountain front retreats by erosion and becomes steadily more embayed, less steep, and lower. Wide fans or pediments coated with alluvium develop meantime, and often the location of the fault cannot then be determined.

Various faults throughout southern California are so old and have been so long inactive that they have little or no direct topographic influence. Some of these are, at least in part, marked by fault-line valleys, as already indicated.

GEOLOGY ALONG THE SOUTHERN MARGIN OF THE ABSAROKA RANGE: A REPLY

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A number of points in the discussion by C. Max Bauer[1] of a recent publication by the writer[2] warrant a reply. Bauer's criticism[3] of the writer's detailed treatment of the subject matter is, of course, a matter of personal opinion. In this, as in other regional studies, it seemed advisable to consider the careful work of other students of geology in adjacent areas, particularly since the conclusions of these workers seemed to be pertinent to this area. Bauer comments[4] that the writer gave no proof, for the statement that the Laramide revolution did not affect all the mountain ranges simultaneously and remarks that the writer had "better . . . . confine his discussions to the contribution which his area makes to a better understanding of the Middle Rockies themselves." The findings of numerous workers on the Laramide revolution were cited by the writer[5] to show that there were several pulsations of the Laramide revolution in Utah, Colorado, Wyoming, Montana, and the Dakotas. These investigations and the conclusions derived from them constitute proof that cannot be ignored. The writer did not confine his discussion to his area alone because if he had, he would have been unable to compare the sequence of events in adjacent mountain ranges with that at the southern end of the Absaroka Range. Without this comparison much of the significance of the orogenic events so well shown in the Absaroka region would be lost.

Bauer[6] believes that the most widespread Tertiary plain of erosion

[1] "Discussion: Geology along the Southern Margin of the Absaroka Range, Wyoming," *Jour. Geol.*, Vol. XLVIII (1940), pp. 324–27.

[2] "Geology along the Southern Margin of the Absaroka Range, Wyoming," *Geol. Soc. Amer. Spec. Paper No. 20* (1939).

[3] *Op. cit.*, p. 324.

[4] *Ibid.*

[5] *Op. cit.*, pp. 104–6.

[6] *Op. cit.*, p. 325.

and deposition was developed in late Eocene time; that epeirogenic uplift came after deposition of the Oligocene strata; that the Laramide revolution closed with the Oligocene; and that ". . . . for central Wyoming and for the Middle Rocky Mountains the close of Laramide deformation can be clearly defined by an understanding of the character of forces effecting the deformation."[7]

The writer, on the other hand, believes that the most widespread Tertiary plain of erosion and deposition was developed after early Pliocene time; that the epeirogenic uplift occurred probably in late Pliocene time; and that, far from being clearly defined, the close of the Laramide revolution must be arbitrarily chosen. It is true that there were two types of movements during Tertiary times in this region: first, orogenic, and second, epeirogenic; but if these types are used as criteria for determining the close of the Laramide revolution, the time of regional peneplanation or high-level pedimentation and the time of subsequent epeirogenic uplift must first be established.

A number of field relationships near the southern margin of the Absaroka Range make Bauer's hypothesis that this widespread surface developed during late Eocene time most unlikely:

1. Upper Eocene strata were deposited against the flank of the Wind River Mountains, across the Washakie Range, and on the Owl Creek Mountains, invariably on a surface of considerable relief. The overlying Oligocene strata were likewise deposited on a surface of high relief and overlie strata ranging in age from pre-Cambrian to Upper Eocene. These relationships may be observed on the stream divide 3 miles southwest of Mount Kent, where the Wiggins formation rests on a rugged surface of Cambrian and Mississippian strata. There is no suggestion of a late Eocene plain here. A similar relationship exists all along the Laramie Mountains of southeastern Wyoming, where the Oligocene rocks were deposited on a surface of high relief which gives no suggestion of late Eocene planation. The rugged topography of the pre-Cambrian core of the Granite Ranges of central Wyoming, on which the Oligocene strata were deposited between Tin Cup Mountain and Independence Rock, does not suggest widespread late Eocene planation of the harder

[7] *Ibid.*

rocks. W. J. Sinclair and Walter Granger[8] state that after the close of Eocene sedimentation a fairly mature topography developed before the beginning of Oligocene deposition on the Sweetwater escarpment. Bauer, on the other hand, states that this "is a very regular surface" and part of the Wind River plain.

2. The high-level Wind River plain, which is cut in hard pre-Cambrian rocks near Gannett Peak in the Wind River Mountains, is clearly visible from the southern margin of the Absaroka Range. Twenty-five miles to the north and northeast of this plain are remnants of a similar widespread surface at nearly the same elevation, cut in Wiggins strata.[9] Nevertheless, Bauer states: "Immediately to the north of the Wind River Range no comparable plain is now traceable because of volcanic disturbance."[10] In the final analysis, any correlation of remnants of erosion surfaces is a matter of interpretation, but anyone who views both the Wind River plain and the Absaroka plain in the same panorama from either Crow Mountain or Wiggins Peak will be impressed by their striking similarity of character and elevation.[11] The writer has tentatively correlated them for these reasons and because of the following point.

3. The Wind River plain would have to be warped down sharply to the north and northeast in order to pass below the Wiggins strata, which are younger than Eocene, and there is no evidence that it does. On the other hand, it would merely have to maintain a comparable elevation to correlate directly with the surface remnants on the Absaroka Range to the north. Even if the plain developed on the Absaroka volcanic rocks did not correlate with the Wind River surface, the mere fact that it cuts Oligocene(?) and possibly younger rocks indicates quiescent conditions here at a time when Bauer believes there was epeirogenic uplift and the beginning of regional degradation.

4. As stated before,[12] it must have taken a long time for such a

[8] "Eocene and Oligocene of the Wind River and Bighorn Basins," *Bull. Amer. Mus. Nat. Hist.*, Vol. XXX (1911), p. 99.

[9] Love, *op. cit.*, pp. 111–12 and Pl. 2.

[10] *Op. cit.*, p. 325.

[11] See also Eliot Blackwelder, "Post-Cretaceous History of the Mountains of Central Western Wyoming" *Jour. Geol.*, Vol. XXIII (1915).

[12] Love, *op. cit.*, p. 113.

surface as the Wind River plain to develop on hard pre-Cambrian and younger rocks. Bauer[13] attempts to explain this by stating: "Moreover, Eocene time was fully half that of the entire Tertiary, and therefore ample time had elasped since the earlier deformations to develop this plain." This time estimate is contrary to data obtained by radioactive determinations, contrary to estimates based on evolutionary trends shown by faunas and floras, and contrary to estimates made by studies of sedimentary and diastrophic history.

The epeirogenic uplift could not have occurred before the high-level erosion and deposition surface was formed. Bauer[14] states that this uplift probably began in early Miocene time and cites as evidence (1) the arid types of fauna and flora and (2) the beginning of general stream degradation. An objection to the first point is that the Miocene faunas and floras of this entire region are very imperfectly known and are badly in need of revision. For example, much of the Yellowstone National Park flora formerly considered to be Miocene now appears to be Eocene and Oligocene. Hence, in our present state of knowledge, any conclusions formed on the basis of known faunas and floras must be regarded as tentative.

Concerning the second point, if general stream degradation began in early Miocene time, it would be hard to account for the following relationships, all of which were discussed in the writer's paper:[15]

1. Middle and Upper Miocene strata were deposited around rugged granite knobs along the Sweetwater River on the crest of the Granite Mountains.

2. Similar strata were deposited on the Oligocene across the buried southern end of the Wind River Mountains.

3. Miocene strata are present on the crest of the Laramie Range.

4. The Sierra Madre and Medicine Bow Mountains of southeastern Wyoming and northern Colorado were largely buried by Pliocene strata.

All this deposition of strata on the crests of buried mountain ranges certainly does not suggest regional degradation and is likewise incompatible with Bauer's belief that the course of Wind River was established during Miocene time.

[13] *Op. cit.*, p. 326. [14] *Ibid.* [15] Love, *op. cit.*, pp. 111–15.

Bauer[16] does not evaluate the writer's reasons for rejecting the idea of a late Eocene plain other than to say that the writer's conclusions based on the geographic position and vertebrate fauna of the Middle and Upper Miocene strata at Split Rock are unwarranted because the formation has never been named or described, and its areal extent has not been mapped. The writer again wishes to point out that both Joseph Leidy[17] and M. R. Thorpe[18] have described Miocene fossils from tuff beds in this locality and that these beds are on the crest of the partially buried Granite Mountains. These facts are all that are fundamentally necessary to form a basis for the writer's previously stated conclusions, which in this case are not affected by either the name, the description, or the areal distribution of the strata. These beds and their stratigraphic relationships can be observed in the Sweetwater Valley near Split Rock, and the identified fossils are still in existence.[19]

It is well established that from time to time between the Eocene and Pleistocene there were both compressional and tensional localized movements of various intensities in this region.[20] An extreme example of this is in southeastern Wyoming where the Laramide movements on the east side of the Laramie Range terminated in Eocene time, whereas 70 miles to the west, in the Saratoga Basin, considerable differential movement continued well into Pliocene time. Bauer[21] explains such instances as these by stating that "there were later local folds and possibly some reverse faulting, but this deformation followed the normal faulting of the Miocene and is

[16] *Op. cit.*, p. 327.

[17] "Remarks on a Collection of Fossils from the Western Territories," *Phil. Acad. Nat. Sci., Proc.*, Vol. XXII (1870), pp. 109–10.

[18] "The *Merycoidodontidae:* An Extinct Group of Ruminant Mammals," *Peabody Mus. Nat. Hist., Mem.*, Vol. III, Part IV (1937).

[19] For their present location see Thorpe, *op. cit.*

[20] Love, *op. cit.*, p. 114; R. L. Nace, "Summary of the Late Cretaceous and Early Tertiary Stratigraphy of Wyoming," *Wyo. Geol. Surv. Bull. 26* (1936), and "Geology of the Northwest Part of the Red Desert, Sweetwater and Fremont Counties, Wyoming," *Wyo. Geol. Surv. Bull. 27* (1939); C. J. Hares and H. J. Cook, "Rocky Revolution" (abstr.), *Geol. Soc. Amer. Proc., 1936* (1937), pp. 77–78.

[21] *Op. cit.*, p. 325.

assigned to late Tertiary." This explanation is unsound, because no worker has ever produced detailed structural, stratigraphic, and paleontologic evidence to show that most of the normal faulting in this region is confined to the Miocene. Consequently, at the present time, there is no basis for separating the later Tertiary localized compressional movements from the earlier movements of the same type.

If, then, as the writer believes, available evidence indicates that completion of the widespread plain and subsequent epierogenic uplift occurred during or at the close of Pliocene time, the burden of proof is certainly on those who believe that "the close of Laramide deformation can be clearly defined" rather than arbitrarily chosen.

Bauer[22] also remarks that: "Contrary to Love's statement, both Big Horn Basin and Wind River Basin had been so filled with sediment by the close of the Eocene that only the higher peaks of the Wind River, Granite, Owl Creek, and Big Horn ranges protruded above the plain." Actually, the writer stated[23] that the Washakie and Owl Creek ranges were largely buried by the close of Eocene time. However, as the writer has already pointed out,[24] the positions of erosion remnants of Upper Eocene and Oligocene(?) strata on the southern end and crest of the Big Horn Mountains[25] and on the northwestern end and southern end of the Wind River Mountains indicate that considerably more than "only the higher peaks [of these ranges] . . . . projected above the plain" at the close of Eocene time. The interpretation of time of filling of the Wind River and Big Horn basins depends fundamentally, of course, upon whether one accepts Bauer's view of late Eocene planation or the writer's view of Pliocene planation.

[22] *Ibid.*, p. 327.

[23] Love, *op. cit.*, pp. 110–11.

[24] *Ibid.*, pp. 78, 84, 111.

[25] N. H. Darton, "Geology of the Bighorn Mountains," *U.S. Geol. Surv. Prof. Paper 51* (1906), pp. 67–70.

REVIEWS

The Origin of Submarine Canyons: A Critical Review of Hypotheses. By DOUGLAS JOHNSON. ("Columbia Geomorphic Studies," No. 3.) New York: Columbia University Press, 1939. Pp. viii+126. $2.50.

The Origin of Submarine Canyons is a reprinting of a series of papers that appeared earlier in the *Journal of Geomorphology*. It reviews the hypotheses that have been advanced by various writers to account for the canyons and presents the author's hypothesis of sapping by submarine springs. An extensive Bibliography, including the earlier literature, is a particularly valuable feature of the book.

Ruling out the possibility of a tectonic origin for the canyons, the author treats the remaining hypotheses under three headings: "Hypotheses of Subaerial Origin," "Hypotheses of Submarine Origin," and "Hypotheses of Subterranean Origin." A subaerial origin is considered unlikely because of the enormous vertical oscillations of sea-level demanded and the apparent lack of independent evidence for such oscillations. Hypotheses of submarine origin include submarine landsliding and submarine currents of which turbidity currents as discussed by Daly are given most attention. After an extended discussion it is concluded that the hypothesis of turbidity currents does not satisfactorily meet the conditions.

To the reviewer it seems that the author has devoted an undue amount of space (most of pp. 35–57) to turbidity currents in fresh water—the Rhone delta, and fresh river water which, in spite of its turbidity, often floats on sea water—because, though interesting, the behavior of muddy fresh water seems beside the point as argument unfavorable to Daly's concept of turbidity currents in salt water caused by wave-stirring of sediments on the shelves. The space given to critically pertinent evidence against the turbidity hypothesis is small, and the treatment (esp. p. 57) raises doubt whether the author's concept of the implications of the turbidity-current hypothesis is sufficiently close to Daly's to make the criticisms pertinent. The author seems to underestimate the self-accelerative ability of a turbidity current once started on a slope veneered with recently deposited fine sediment. Yet, curiously enough, in a later part of the book he gives this factor (in a modified form as periodic mud flows)

considerable weight in keeping open, and perhaps enlarging, canyons caused by subterranean spring sapping.

Under hypotheses of subterranean origin such possibilities as subterranean river outlets, solution along faults, or nondeposition along faults due to up-welling waters are discussed briefly and dismissed in favor of the author's hypothesis of sapping by submarine springs.

The essentials of that hypothesis may be briefly summarized as follows: (1) Stratigraphic conditions on the continental shelves and adjacent coastal plains are or have been favorable to strong artesian circulation which, reinforced by water squeezed out of the sediments by compaction, may have given rise to powerful submarine springs on the continental slope capable of producing canyons by headward sapping. (2) Inasmuch as considerable canyons have been produced by subaerial spring sapping—for example, in the Grand Canyon region, the lava plateaus of Idaho, and the limestone plateau of Florida—somewhat similar features might be expected where springs emerge beneath the sea. (3) Emerging beneath the sea, artesian fresh water would be expected to rise, and there would be no stream to carry away the debris. So it is not, at first thought, evident why a canyon should be produced. Solution seems to be the answer. All rocks, even the sands and clays that are supposed to constitute the bulk of the shelf sediments, are somewhat soluble, especially in carbonated water, therefore, ". . . . *given sufficient* time the longest and deepest canyon in the most insoluble rocks might be produced by solution alone." (4) In earlier stages numerous springs started canyons on the continental slope, but eventually certain larger springs gained advantage and developed the major canyons, cutting off the water supply from adjoining canyons and ravines. (5) Sediment accumulating on the continental slope periodically becomes unstable and, as mud flows, slides into and down the canyons, thus removing any debris that may have been carried out by the springs and accumulated around their mouths and keeping clear or possibly enlarging canyons and ravines on the continental slope that have been abandoned by the springs that made them.

Such is the essence of the spring-sapping hypothesis. Geologists will find much to question in the author's concept of continental-shelf structure, particularly where, as in Figure 1, coastal-plain dips are projected as straight lines to outcrop on the continental slope, and will find it difficult to avoid the feeling that the more probable features of shelf structure have been minimized and less probable structures, favorable to the postulated artesian circulation, have been unduly stressed. Also, they may well question the former position of outcrops of the lower coastal-plain aquifers

". . . . some thousands of feet above sea-level throughout much of the long Schooley cycle."

Geomorphologists will note that most of the submarine canyons and ravines as depicted in the three charts published in the book and as shown on the larger Veatch and Smith charts fail to display one of the most characteristic features of spring sapping—steep, amphitheater-like heads, or even the half-cone-shaped heads that might be expected in unconsolidated rock.

Even granting solution as postulated by the author, it is difficult to see why a canyon should be formed instead of a funnel-shaped pit where low-density fresh water from a spring starts rising toward the surface of the ocean.

Finally, both geomorphologists and geologists will find difficulty in understanding why the locus of a submarine spring, as it migrates backward into the continental slope, should increase steadily in absolute elevation so as to produce the rather evenly sloping canyon and ravine bottoms which the maps reveal, or why all the springs should have started at the bottom of the continental slope and have migrated upward and backward essentially parallel to the surface of the slope to what one must infer is their present position at the heads of the canyons some thousands of feet higher.

Last, the reader may well question whether the mud flows—in effect, self-accelerating turbidity-density currents—which the author invokes to keep the spring-formed canyons clear may not play a much more important role in the actual canyon-cutting than the author suspects.

Irrespective of the merit of the spring-sapping hypothesis, all students of the submarine-canyon problem must be grateful to the author for assembling in readily available form a critical summary and bibliography of the literature, for it is widely scattered and some of it dates back nearly a hundred years. Also, occasional critical reviews of the various hypotheses that have been offered to explain the canyons is an effective stimulus to further research.

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